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LATE QUATERNARY GLACIAL HISTORIES AND HOLOCENE PALEOENVIRONMENTAL RECORDS FROM NORTHEAST AND SOUTHWEST ELLESMERE ISLAND, NUNAVUT, CANADA

by

IAIN RODERICK SMITH ©

A thesis submitted to the Faculty of Graduate Studies and Research
in partial fulfillment of the requirements for the degree of Doctor of Philosophy

Department of Earth & Atmospheric Sciences

Edmonton, Alberta

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FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled LATE QUATERNARY GLACIAL HISTORIES AND HOLOCENE PALEOENVIRONMENTAL RECORDS FROM NORTHEAST AND SOUTHWEST ELLESMERE ISLAND, NUNAVUT, CANADA submitted by IAIN RODERICK SMITH in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY.

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ABSTRACT

Ice-free areas beyond the limits of the last glaciation have been proposed for much of Ellesmere Island. This hypothesis is addressed by reconstructing the late Quaternary glacial history of Lake Hazen Basin and eastern Hazen Plateau, and by coring extant lake basins beyond proposed ice margins. Diatom records from these lakes, and stable isotope records from emergent basins on Hoved Island, southwest Ellesmere Island, were used to assess Holocene environmental changes in the High Arctic.

Lateral meltwater channels, moraines and other geomorphic evidence indicate that a large trunk glacier emanating from the Grant Land Mountains coalesced with Agassiz and Greenland ice, inundating Hazen Plateau. The configuration of deglacial margins related to the trunk glacier, Holocene ice-contact deltas in Robeson Channel, and cosmogenic ^{36}Cl dating of erratics, indicate that this occurred during the last glaciation. Thus, ice-free regions did not exist in the Lake Hazen region.

Breakup of marine-based ice margins between 9 and 8 ka BP led to a retreat of Grant Land Mountain trunk ice, and deglaciation of outermost Hazen Plateau. Plateau ice caps, however, persisted and expanded over highland regions in early Holocene. Between 7 and 6 ka BP, ice retreated to the heads of regional fiords and valleys, after which it remained stable. Breakup of ice within the proto-Lake Hazen basin occurred between 5.3 and 5 ka BP, at which point Grant Land Mountain ice had retreated to near its modern limits.

Diatom abundances serve as a proxy record of summer lake ice cover. Results from the Lake Hazen region indicate a gradual climatic amelioration between 5 and 4 ka BP. The greatest reduction in ice cover occurred between 4 and 3 ka BP, after which, a cooling led

to a decline in diatom abundances in higher elevation lakes, while those lower down maintained high levels until ~2 ka BP.

Basal diamicts in cores from two lakes on Hoved Island records the retreat of formerly grounded ice in central Baumann Fiord ~9.3 ka BP. Isotopic records of *Lobatula lobatulus* detail a stepped ice retreat during early to mid-Holocene, beyond the resolution of previous studies of postglacial emergence.

This thesis is dedicated the memory of Kent Holden

A superb friend and colleague
whose potential will never be fully realized
and who's wit, insight and contribution are dearly missed.

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Throughout the course of my Ph.D. studies I have experienced tremendous support, encouragement and inspiration from a number of individuals. Though I am the author, an undertaking such as this dissertation can never be considered solely my own efforts.

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Colleagues and friends have also made important contributions through the years. From preserving my sanity, to acts of purposely trying to lose it, they have inspired, cajoled and assisted me through many stages of my Ph.D. I particularly wish to thank the following individuals for discussing, arguing and bashing ideas out over coffee, that have helped to focus my thoughts: Trevor Bell, Kent Holden, James Hooper, Scott Lamoureux and Colm O'Cofaigh. In addition to the many savage reviews of manuscripts provided me by John England, this dissertation has benefitted from the reviews and discussion of various manuscripts by Dr.'s Martin Sharp and John Shaw. Martin Sharp has shown himself to be a valued and trusted sounding board, and I have greatly enjoyed the interaction with him through the years (though not his choice of where to dig snowpits!).

Technical assistance was generously provided me by: Alwynne Beaudoin, Provincial Museum of Alberta, who identified seeds and cladocera egg cases; Denis Delorme (Research Scientist Emeritus), National Water Research Institute, Canada Centre for Inland Waters, who identified the ostracodes; Catherine LaFarge-England, Department of Biology, Duke University, who identified bryophyte samples submitted for dating; Randy Pakan, Department of Earth & Atmospheric Sciences, University of Alberta, who assisted with photographic and image processing; Ian Walker, Department of Biology, Okanagan University College, who identified the chironomid head capsules. Dr. Karlis Muehlenbachs, Department of Earth & Atmospheric Sciences, University of Alberta is thanked for affording me the opportunity of conducting the stable isotope research as part of a graduate research project.

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TABLE OF CONTENTS

1. INTRODUCTION 1
    1.1 Outline of Dissertation 2
    1.2 References 7

2. DIATOM-BASED HOLOCENE PALEOENVIRONMENTAL RECORDS FROM THE LAKE HAZEN REGION, NORTHERN ELLESMERE ISLAND, NUNAVUT, CANADA 11
    2.1 Introduction 12
        2.1.1 Study Area 13
        2.1.2 Limnology 15
    2.2 Methods 16
        2.2.1 Sediment Cores 16
        2.2.2 Diatoms 16
        2.2.3 Radiocarbon Dating 20
    2.3 Results 22
        2.3.1 Lake Sediments 22
        2.3.2 Diatom Stratigraphies 23
            2.3.2.1 Appleby Lake 27
            2.3.2.2 Brainard Lake 28
            2.3.2.3 Hart Lake 29
            2.3.2.4 Whisler Lake 30
    2.4 Discussion 31
        2.4.1 Nature of the Diatom Records 31
        2.4.2 Ice-cover Proxy Record 33
        2.4.3 Provisional Paleoenvironmental Reconstructions 36
        2.4.4 Comparisons with other Regional Paleoenvironmental Records 40
            2.4.4.1 Early to Mid-Holocene Records 41
            2.4.4.2 Post 3 ka BP and the Neoglacial 43
    2.5 Conclusions 45
    2.6 References 47
    2.7 Appendix I 60

3. INTERPRETATION OF DIAMICTIC SEDIMENTS WITHIN HIGH ARCTIC LACUSTRINE CORES: EVIDENCE FOR LAKE ICE-RAFTED DEPOSITION 62
    3.1 Introduction 63
        3.1.1 Study Area 65
        3.1.2 Core Sedimentology 67
    3.2 Methods 71

3.2.1 Sediment Cores 71
3.2.2 Clast Fabric Analysis 71
3.2.3 Particle-size 72
3.2.4 Subfossil Remains 72
3.2.5 Ostracodes 73
3.2.6 Diatoms and Chrysophyte Stomatocysts 73
3.2.7 Chironomid Head Capsules 74
3.3 Results 74
3.3.1 Clast Fabrics 74
3.3.2 Particle-size 78
3.3.3 Ostracodes 81
3.3.4 Diatoms and Chrysophyte Stomatocysts 85
3.3.5 Chironomid Head Capsules 87
3.4 Discussion 87
3.4.1 Deglacial History and Diamict Chronologies 87
3.4.2 Tills, Flow or Ice-Rafted Deposits? 92
3.4.3 Ice-Rafting 94
3.4.4 A Possible Mechanism for Ice-Rafting 97
3.5 Conclusions 100
3.6 References 102
3.7 Appendix 109
Appendix 3.1 109

4. TERTIARY OUTLIERS AND REGIONAL LANDSCAPE EVOLUTION OF EASTERN HAZEN PLATEAU, NORTHERN ELLESMERE ISLAND, NUNAVUT, CANADA 111
4.1 Introduction 112
4.1.1 Regional Geology 112
4.2 Field Evidence 115
4.2.1 Hazen Plateau 115
4.2.2 Tertiary Outliers 117
4.3 Discussion 119
4.3.1 Regional Valley and Fiord Formation 119
4.4 Conclusions 120
4.5 References 122

5. THE LATE QUATERNARY GLACIAL HISTORY OF LAKE HAZEN BASIN AND EASTERN HAZEN PLATEAU, NORTHERN ELLESMERE ISLAND, NUNAVUT, CANADA 124
5.1 Introduction 125
5.1.1 Previous Investigations 125
5.1.1.1 Recent Revisions of the Quaternary History 127
5.2 Study Area 128
5.2.1 Physiography and Climate 128

5.3 Research Methods 129
5.4 Field Observations 130
    5.4.1 Glacial Geomorphology 130
5.5 Glacial Reconstruction 142
    5.5.1 Ice Buildup to the Last Glacial Maximum 142
    5.5.2 Initial Deglaciation of Hazen Plateau 144
    5.5.3 Marine Incursion 149
    5.5.4 Plateau Ice Caps 151
    5.5.5 Retreat of Ice Inland From Hazen Plateau 152
5.6 Discussion and Conclusions 156
    5.6.1 Last Glacial Maximum 156
    5.6.2 Deglaciation of Robeson Channel and Northeast Hazen Plateau 159
    5.6.3 Final Deglaciation 162
5.7 References 164

6. THE DEGLACIAL AND HOLOCENE STABLE ISOTOPE STRATIGRAPHY OF THE FORAMINIFERA *Lobatula lobatulus*, FROM AN ISOLATION BASIN, HOVED ISLAND, CANADIAN HIGH ARCTIC 171
6.1 Introduction 172
    6.1.1 Study Area 174
    6.1.2 Physiography 174
    6.1.3 Glacial History 174
    6.1.4 Oceanography 175
6.2 Methods 179
    6.2.1 Sediment Cores 179
    6.2.2 Radiocarbon Dating 179
    6.2.3 Stable Isotope Analysis 179
6.3 Results and Discussion 187
    6.3.1 Core Sedimentology 187
    6.3.2 Stable Isotopes 188
        6.3.2.1 Vital Effects and Sample Variability 188
        6.3.2.2 *Lobatula lobatulus* 190
        6.3.2.3 *Portlandia arctica* 196
6.4 Conclusions 198
6.5 References 200

7. SUMMARY 208
7.1 References 214

LIST OF TABLES

2.1 Morphometric parameters and chemistry of the study lakes 18
2.2 Radiocarbon date list 21
3.1 Morphometric parameters and other characteristics of the study lakes 68
3.2 Abundances of subfossils in Appleby, Connell and Brainard lake sediments . . . 83
3.3 Ostracode species composition (% of total) for Connell and Brainard lakes 84
3.4 Abundance and species composition of chironomid head capsules 88
3.5 Radiocarbon date list and sample data 90
5.1 Radiocarbon date list . . . . 131 - 132
5.2 Cosmogenic ^{36}Cl date list 133
5.3 Oxygen and deuterium isotope analyses of buried ice 141
6.1 Radiocarbon date list 183
6.2 Stable isotope results for replicate samples of *Lobatula lobatulus* 185

LIST OF FIGURES

1.1 Location maps of the two principal study areas on Ellesmere Island 3
2.1 Location map of study area on northeastern Ellesmere Island 14
2.2 Topographic map of field area showing location of the lakes cored 17
2.3 Frequency diagram of the diatom assemblage from Appleby Lake 24
2.4 Frequency diagram of the diatom assemblage from Brainard Lake 25
2.5 Frequency diagrams of the diatom assemblages from Hart and Whisler lakes ... 26
2.6 Provisional paleoenvironmental reconstruction 34
2.7 Oblique air photograph looking west across Hazen Plateau and Lake Hazen ... 39
3.1 Study area located on northeastern Ellesmere Island 64
3.2 Topographic map of the region and study lake bathymetry and coring sites 66
3.3 Sediment core logs of the two facies assemblages 69
3.4 Photographs of section of the gravel-rich diamict from three lakes 70
3.5 Schmidt equal-area stereo plots of clast fabrics from gravel-rich diamicts 75
3.6 Eigenvalue S_1 vs. S_3 plot of the gravel-rich diamicts 76
3.7 General shape triangle plot of the clast fabric eigenvalue data 77
3.8 Photograph of the lowermost section of Whisler core #1 79
3.9 Particle-size diagram 80
3.10 Photos and scanning electron micrographs of subfossil remains 82
3.11 Diatom % abundance for Whisler core #1 86
3.12 Deglacial ice margins for 7.5-7 and 6.5-6 ka BP 89
3.13 Age vs. depth relationships in Appleby, Brainard and Whisler lakes 93
3.14 Oblique photograph of the kame deltas adjacent Whisler Lake 96
3.15 Graph showing cummulative depth (%) of Connell, Hart and Whisler lakes 99
4.1 Maps showing distribution of Tertiary and Mesozoic strata in study area 113
4.2 Region of Tertiary outliers south of Lake Hazen 114
4.3 Cross-section showing position of different strata along Salor Creek 116
5.1 Location map of the study area 126
5.2 Prominent glacial erosional features 134
5.3 Oblique air photograph of the Chandler Fiord region 135
5.4 Prominent glacial erosional and depositional landforms 136
5.5 Oblique air photograph looking southeast over Hazen Plateau 138
5.6 Oblique air photograph looking northwest across Lake Hazen Basin 140
5.7 Fossiliferous erratics found above Eastwind Bay 143
5.8 Oblique air photograph looking west up Conybeare Fiord 147
5.9 Paleogeographic maps of ice margins at 8, 7, 6 and 5 ka BP 148
5.10 Ice margins during the LGM and at 10 and 9 ka BP .... 157 - 158
6.1 Study area on southwest Ellesmere Island 173
6.2 Circulation patterns within the Arctic Archipelago 178
6.3 Oblique photograph of the study lakes 180
6.4 Sediment logs and spatial distribution of the three study lakes 181
6.5 Photos of the lower core sections from Walton and Izaak lakes 182
6.6 Graph of the concentration of Lobatula lobatulus tests in the Walton core 186

6.7 Oxygen and carbon isotope stratigraphy from Walton Lake core 189
6.8 Stable isotope stratigraphy calibrated in ^{14}C years BP 191
6.9 Isotope data of *Portlandia arctica* samples 197

LIST OF APPENDICES

2.1 Diatom species list and taxonomic authority 60
3.1 Diatom and chrysophyte assemblages from diamict samples 109

INTRODUCTION

1.1 OUTLINE OF DISSERTATION

This dissertation is a multi-disciplinary study of late-Quaternary glacial history and Holocene paleoenvironmental change within northeast and southwest Ellesmere Island, Canadian High Arctic (Fig. 1.1). Results are presented and discussed in five papers.

The main study area focuses on the continental interior region of Lake Hazen Basin and eastern Hazen Plateau, northeastern Ellesmere Island (Fig. 1.1). The original objectives of this project were proposed in light of previous glacial reconstructions by England (1976, 1978, 1983), Bednarski (1986) and Retelle (1986) who had worked around the adjoining coastal regions of northeastern Ellesmere Island. This study was thus designed to test these proposed models of past glaciations, specifically addressing questions with regards to the existence of ice-free regions during the Last Glacial Maximum (LGM). Coring of extant lakes beyond the proposed maximum glacial margins (Fig. 1.1) was hoped to reveal paleoenvironmental records spanning the last glaciation. This would provide a unique record, and potentially resolve some of the questions surrounding the buildup of ice in a region where glaciers remain greatly constrained by aridity (Bradley & England 1978; Alt 1985; Alley *et al.* 1993).

Chapter 2 examines the diatom-based Holocene paleoenvironmental records from 8 shallow lake basins (4 of which were studied in detail) within the Lake Hazen region. In the Canadian High Arctic, paleoenvironmental records have mostly been derived from cores of the various ice caps (cf. Koerner 1979), or from the marine environment, utilizing various proxy records (e.g., temperature, salinity or sea ice abundance; cf. Aksu 1985; Dyke *et al.* 1997). There have been relatively few terrestrially-based paleoenvironmental records, amongst the most important of which have been studies of lake sediments (Bradley 1990). Lakes offer the advantage of occurring throughout the topoclimatically diverse Arctic region, and contain a number of autochthonous and allochthonous proxy environmental indicators (cf. Smol & Douglas 1996). Of these, diatoms have become the most widely used, and permit the identification of changes in the biotic environment (pH, habitat, nutrient fluxes) as well as the physical lake environment (sediment input and the persistence of a summer ice floe; Smol 1988). Paleoenvironmental records constructed from this study indicate

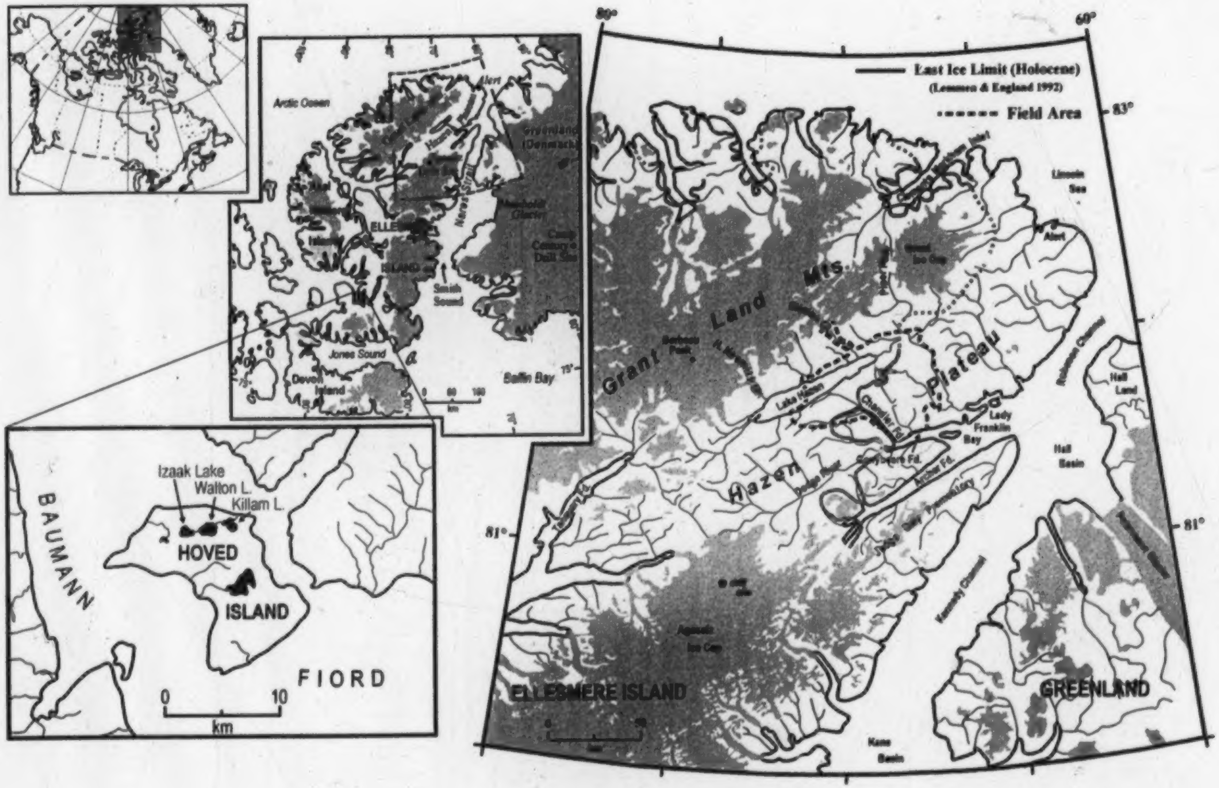
Last Ice Limit (Holocene)
(Lemmen & England 1992)
Field Area
80°
60°
83°
81°
81°
Grant Land Mts.
Hazen Plateau
Lake Hazen
Lady Franklin Bay
Hall Basin
Hall Land
Lincoln Sea
Alert
Kane Basin
Agassiz Ice Cap
ELLESMERE ISLAND
GREENLAND
Arctic Ocean
Greenland (Denmark)
Nares Strait
Smith Sound
Jones Sound
Devon Island
Baffin Bay
ELLESMERE ISLAND
Izaak Lake
Walton L.
Killam L.
HOVED
ISLAND
BAUMANN
FIORD
0
10
km

Figure 1.1 Location maps of the two principal study areas on Ellesmere Island, Canadian Arctic Archipelago. The Lake Hazen region is located on northeastern Ellesmere Island (rightmost figure). Hoved Island is located in southwest Ellesmere Island (lower left figure). Grey shading on the centre and rightmost figure depicts contemporary glacial cover.

considerable changes have occurred through the mid to late-Holocene. Variability in the records highlights strong local topoclimatic differences, and also contrast with records derived from coastal regions of Ellesmere Island.

Chapter 3 examines the morphology and possible genesis of basal diamictic facies within 6 of 16 lake basins cored in the Lake Hazen region. In the other 10 basins, basal sediments comprised a mixed facies of sorted sediment, rhythmites and/or short sections of diamict with sand and gravel interbeds. The later facies appears consistent with deposition in proglacial lakes. The most direct interpretation of the diamict units from cores devoid of sorted sediment, is that they represent tills, and therefore could be used as an important constraint on the regional deglacial history. This study thus examines the diamict sedimentology, clast fabric and subfossil remains (diatoms, ostracodes and chironomid head capsules), results of which contradict a till genesis. Further, the absence of sorted sediment in diamict units up to 2.1 m in length suggests they may not be proglacial lake deposits. An alternative model is proposed invoking adfreezing of sediment and rafting by the lake ice cover.

Chapter 4 discusses landscape evolution and formation of the regional valleys and fiords in the Lake Hazen region, including a consideration of what role glacial excavation may have played. It also provides a reinterpretation of the extent of Tertiary erosional remnants south of Lake Hazen. Previously, it has been proposed that the inter-island channels in the Arctic Archipelago were formed by Tertiary fluvial systems, that subsequently became glacially over deepened (Fortier & Morley 1956; Pelletier 1966; Trettin 1991). From this, Denton and Hughes (1981) concluded that glacial erosion is responsible for the major elements of the Canadian Arctic landscape. Within the Lake Hazen region, Miall (1982, 1991) has made similar arguments for the evolution of the large valleys, and fiords. Stellate nodules, a carbonate concretion indicative of cold marine waters (Kemper & Schmitz 1975), were discovered along an exposure in Salor Creek, a major valley crosscutting Hazen Plateau. The presence of these, contradicts the Tertiary designation of the strata in addition to those unconformably overlying the upper Hazen Plateau (Christie

1976; Miall 1979). Instead, Tertiary outliers southeast of Lake Hazen appear to be confined to discontinuous sections in three valley bottoms. This has important ramifications for the timing at which the major valleys formed. It also indicates that post-Eocene excavation (both glacial and fluvial) of these valleys has been minimal, which sharply draws into question the assertion that glaciers excavated the adjoining cliff-bounded fiords which range from 200 - 600 m deep.

Chapter 5 reconstructs the late-Quaternary glacial history of eastern Lake Hazen Basin and Hazen Plateau. This is the first study to document the continental retreat of glaciers from this region, whereas previous investigations have concentrated on coastal and raised marine records. It thus helps to resolve uncertainties about the extent and persistence of continental ice which has allowed researchers to speculate as to its potential contribution to mid to late-Holocene sea level rises (cf. Scott & Collins 1996). The inland retreat of ice was documented by mapping the surficial glacial geomorphology, surveying raised marine and lacustrine sediments, and coring of modern lake basins. Mapping included air photo interpretation of much of Hazen Plateau, and intensive field surveys south and east of Lake Hazen (Fig. 1.1). Chronological constraint is provided by ^{14}C dating of organic material from lake sediments and peat.

This study also addresses concerns regarding the maximum extent of ice during the LGM. England and other researchers working throughout Ellesmere Island have proposed a greatly restricted ice cover, in which glaciers experienced only moderate advances of 5-60 km beyond their present-day margins; the Franklin Ice Complex (reviewed in England 1987; Lemmen & England 1992; England, in press). Contrasting this was the work of Blake and others who instead proposed that glaciers completely inundated Ellesmere Island and most of the Arctic Archipelago, coalescing with the Laurentide Ice Sheet to the south, and with the Greenland Ice Sheet to the east; the Innuitian Ice Sheet (Blake 1970, 1972, 1992). Recent work by Bednarski (in press), Dyke (in press) and England (1998, in press) now hypothesizes that indeed Ellesmere Island and much of the Arctic Archipelago were inundated by ice during the LGM. Thus, this paper no longer addresses the original debate of England and Blake, but instead tests the newly proposed reconstruction of England (1998, in press), in which ice cover at the LGM is considered to have inundated the entire Hazen Plateau. Surficial mapping, ^{14}C dating of raised marine deposits and lake sediments, and ^{36}Cl exposure dating of bedrock and erratics support the interpretation of England (1998, in press) concerning the extent of ice at the LGM. However, arguments are made against the proposed dynamics and geometry of the coalescent Ellesmere-Greenland ice boundary, as well as the timing of ice retreat within the study area.

In Chapter 6, the initial proposal to study High Arctic lacustrine paleoenvironmental records beyond the ice limits at the LGM was extended to include three emergent lake basins on Hoved Island, central Baumann Fiord, southwest Ellesmere Island (Fig. 1.1). These basins lie between 0 and 5.5 m above sea level, well below Holocene marine limit (~100 m), and thus have emerged from the sea only within the past few millennia; therefore, they contain mostly marine sediments. Assessment of the position of these basins with respect to the ice margins at the LGM was based on the work of Hodgson (1985 - model B), who had identified a "drift belt" considered to mark the maximum extent of ice. This falls in accordance with others who have indicated a similarly restricted ice cover during the LGM along western Ellesmere Island (England 1986, 1987, 1990; Sloan 1990; Bell 1992, 1996). Two other models of the last glaciation were also proposed by Hodgson (1985), invoking both a complete regional ice cover (model A; cf. Blake 1970) and one in which ice at the LGM extended an unknown distance down-fiord of the drift belt (model C). The lake basins on Hoved Island were ideally suited to test these hypotheses. They also offered the opportunity to compare and contrast the lacustrine proxy records from the Lake Hazen region, with marine records from the Hoved Island lakes. Stable isotope analyses of foraminifera (δ^{18}O and δ^{13}C) from the marine sediments in these basins is used to elucidate changes in the regional ice cover through the Holocene.

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DIATOM-BASED HOLOCENE PALEOENVIRONMENTAL RECORDS FROM THE LAKE HAZEN REGION, NORTHERN ELLESMERE ISLAND, NUNAVUT, CANADA

2.1 INTRODUCTION

The Arctic is recognized as highly sensitive to Global Climate Change (Roots 1990; Walsh 1991). Consequently, there is growing interest in paleoenvironmental records from this region, including the Paleoclimates of Arctic Lakes and Estuaries program (PALE, Andrews & Brubaker 1994), renewed coring of the Greenland Ice Sheet (GISP2 and GRIP), and the coring of the Penny Ice Cap on Baffin Island (Fisher *et al.* 1996). Ice cores provide the longest records of environmental change, but are limited spatially and by the types of information they retain. Given the topoclimatic diversity (Jacobs & Leung 1981; Maxwell 1981) and expanse of the Canadian Arctic Archipelago (1.3 x 10^6 km^2), other widespread records must be sought. Paleoenvironmental records are further constrained by the fact that during the last glacial maximum (LGM), most of the Archipelago was either covered by ice, or submerged below sea level because of glacioisostasy (Dyke & Prest 1987). Consequently, except for marine-based records, and rare terrestrial sites beyond the limit of ice (i.e., Wolfe 1994), paleoenvironmental records are limited to the Holocene.

Our present understanding of Holocene paleoenvironmental change in the Canadian Arctic has come mainly from marine-based records, including: deglacial sedimentology and postglacial isostatic uplift (England 1992, 1996), the relative abundance of driftwood (Blake 1972; Stewart & England 1983, Dyke *et al.* 1997) and whalebone on raised shorelines (Dyke & Morris 1990; Dyke *et al.* 1996a), and stable isotope stratigraphies of foraminifera (Aksu 1985; Andrews *et al.* 1991). Terrestrially-based paleoenvironmental studies have been relatively few, the most important of which have been conducted on lake sediments (cf. Bradley 1990; Overpeck *et al.* 1997). The widespread occurrence of Arctic lakes makes them ideal study sites. Because of extreme climatic conditions and reduced weathering of terrigenous clastic materials, paleolimnological records from the Arctic may be controlled as much by changes in the physical lake environment (i.e. persistence of a summer ice floe, water temperature, sediment input from glaciers), as they are by biotically sensitive chemical changes (pH and nutrient fluxes).

Several proxy records have been examined within Arctic lakes, including chrysophycean stomatocysts (Duff & Smol 1988), diatoms (Smol 1988), plant macrofossils (Bennike 1995; Fredskild 1995), pollen (Fredskild 1983; Gajewski *et al.* 1995) and sedimentology (Retelle 1986; Lemmen *et al.* 1988; Lamoureux & Bradley 1996). Of these, diatoms have become the most widely used (Blake *et al.* 1992; Douglas & Smol 1993; Douglas *et al.* 1994; Foged 1972, 1977; Smol 1983, 1988; Williams 1990; Wolfe 1991, 1994, 1996a,b; Young & King 1989). Diatoms are a diverse group of algae (class Bacillariophyceae), found within a wide range of aquatic environments. They possess a cell wall (frustule) of biogenic silica, the morphology of which is unique and identifiable to the species level. Uncertainties in taxonomy and ecology hamper some of their usefulness, but this continues to be addressed through autecological studies (Douglas & Smol 1993, 1995; Hamilton *et al.* 1994a; Hickman 1974; Moore 1974).

Most paleolimnological studies in the Canadian Arctic and Greenland have been conducted in coastal localities, whereas the interiors of many large Arctic islands have remained unstudied. This paper examines diatom-based paleoenvironmental records from several lake basins within the continental interior of northern Ellesmere Island (Fig. 2.1). Extant lake basins within and beyond the proposed limit of ice during the LGM (England 1978, 1983) were cored, in order to test this hypothesis, and to compare the local paleoenvironmental reconstructions with others from nearby coastal sites.

2.1.1 Study Area

The area is comprised of three physiographic regions: Grant Land Mountains, Lake Hazen Basin, and Hazen Plateau (Fig. 2.1). The Grant Land Mountains rise abruptly from the north shore of Lake Hazen (157 m above sea level [asl]), reaching 2665 m asl (Barbeau Peak). They support a large icecap, from which trunk glaciers descend to 500-240 m asl above Lake Hazen. Lake Hazen Basin comprises the peripheral lowlands (<300 m asl, ~3500 km^2) surrounding Lake Hazen. Lake Hazen itself is 70 km long, 290 m deep and 540 km^2. Hazen Plateau (25 000 km^2) is a broad, dissected upland, rising to 1300 m asl along its eastern rim where it supports several small plateau icecaps (Hattersley-Smith & Serson 1972). The principal surface rock of Hazen Plateau is the Danish River Formation (Trettin 1994), which consists locally of >620 m of calcareous and dolomitic sandstone and mudrock, tightly folded along a pronounced SW-NE axial plane. A wedge of Upper Paleozoic to Tertiary beds extends northeastward from Lake Hazen for 80 km (Miall 1979). Tertiary

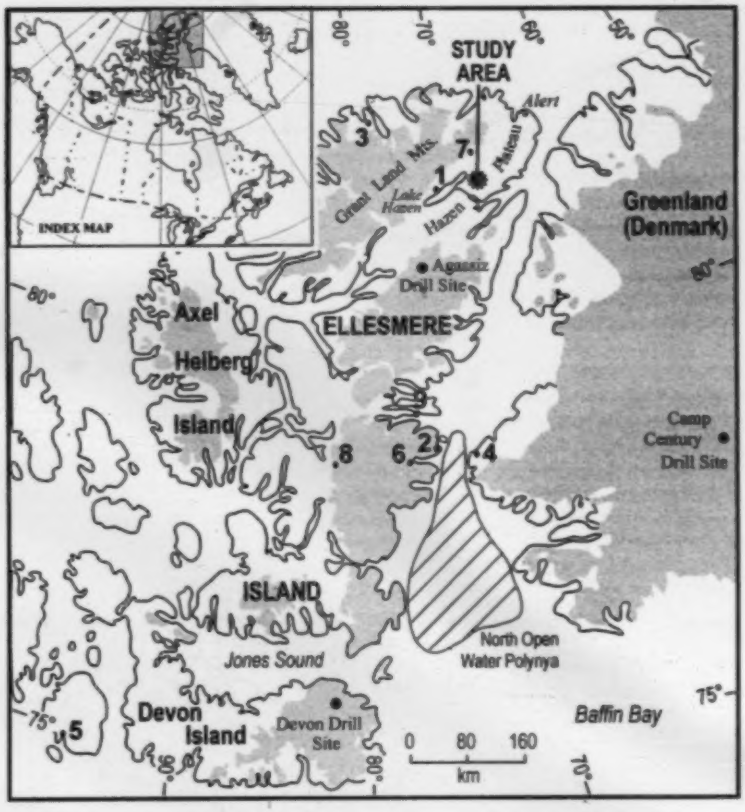

INDEX MAP
80°
70°
60°
50°
STUDY AREA
Alert
3
7
1
Grant Land Mts.
Lake Hazen
Hazen Plateau
Greenland (Denmark)
Agassiz Drill Site
80°
Axel Heiberg Island
ELLESMERE
ISLAND
8
6
2
4
Camp Century Drill Site
North Open Water Polynya
Jones Sound
75°
Devon Island
Devon Drill Site
5
75°
Baffin Bay
0 80 160
km
80°
70°

Figure 2.1 Location map showing the study area on northeastern Ellesmere Island, Canadian High Arctic. Bold numbers refer to sites discussed within the text. Shading depicts contemporary glacier cover.

outliers are also found on Hazen Plateau (Miall 1979).

Climatically, the region constitutes a polar desert (Bovis & Barry 1974), with a mean annual temperature of approximately -20°C, and a mean annual precipitation of <150 mm water equivalent (Thompson 1994). Owing to the continentality and unique topographic characteristics of the region, the area becomes a "thermal oasis" in the brief summer, when temperatures can exceed 18°C. Cold-air drainage and persistent temperature inversions account for mean winter monthly temperatures around -40°C, which are 6 to 10°C colder than Alert on the northeast coast (Thompson 1994; Fig. 2.1).

2.1.2 Limnology

There are numerous small, shallow lakes, and many more permanent and ephemeral ponds throughout Lake Hazen Basin, but are rare on Hazen Plateau. The region is blanketed by a patchy till veneer, and thus most lake basins have bedrock controlled outlets. Other than studies conducted on ponds around Hazen Camp (Oliver & Corbet 1966; Fig. 2.1, site 1) there has been little other physicochemical analysis of lacustrine environments (Hamilton *et al.* 1994b). Lakes are characterised as being ultra-oligotrophic (low in nutrients and productivity), cold monomictic, with pH levels ranging from near-neutral to alkaline (Hamilton *et al.* 1994b; Stewart 1994). Conductance and cation concentrations are strongly related to underlying bedrock. Lakes on the Danish River Fm. are dominated by Ca, and to a lesser degree Mg and Na ions (Hamilton *et al.* 1994b). Maximum lake ice thickness varies from 1.9 to 3.2 m and is typical for High Arctic lakes (Doran 1993; Heron & Woo 1994). There is a strong topoclimatic gradient within this region, and lakes higher on the plateau have greater ice thicknesses (>2.2 m). During 1992 and 1993, lakes in Lake Hazen Basin cleared of ice between mid and late July (excluding Lake Hazen). The snowline never rose above 640 m asl in 1992, thus lakes on Hazen Plateau developed only small peripheral shore leads. Although the summer of 1993 was warmer (temperatures up to 18.5 °C), lakes on Hazen Plateau retained between 70-90% ice cover at the beginning of August. Freeze-up within the High Arctic usually begins by late August to mid-September (Doran 1993; Woo 1980).

2.2 METHODS

2.2.1 Sediment Cores

Sixteen lake basins were cored in the spring of 1993 and 1994. Lakes were selected to test the proposed limit of ice during the LGM (England 1983), and to provide records from a variety of regional and topographic settings. Both a simple percussion corer (Reasoner 1993) and a vibracorer (Smith 1992) were used to extract cores from the deepest basin of each lake. Logistical constraints generally allowed the recovery to only a single core from each lake. Cores were capped and sealed in the field, shipped south and then frozen and split lengthwise with a diamond saw at Core Laboratories, Calgary. The cut face was washed, while frozen, to remove any contamination from the splitting process.

2.2.2 Diatoms

Sediment cores from 8 lakes were initially inspected for diatom remains. In three lakes, *Bender*, *Hodges* and *Linn* (unofficial lake names are presented in italics in the first instance; Fig. 2.2), diatoms were absent or very sparse. Similarly, diatoms were found only in the upper 3 cm of sediment from *Stewart* Lake. Sediments from Appleby, *Brainard*, *Hart* and *Whisler* lakes contained the most complete diatom records, and are thus the focus of this study (Fig. 2.2, Table 2.1).

Small (~3 g) samples were cut every 5 cm from the centre of thawed core sections (core lengths varied from 0.72 to 3.34 m). Each sample included no more than 5 mm vertically of sediment. Because of the high clay and fine-silt content, considerable problems (broken valves, incomplete particle disaggregation) were encountered when wet samples were first oven-dried (105°C) and then digested. To counter this, replicate samples were taken, and the sample for diatom analysis was digested wet, while the other was dried. Wet weights were then adjusted accordingly, and all results are presented on a dry weight basis (diatom valves/gram dry sediment (VGS)).

Samples were pretreated for organic removal by digesting them in concentrated sulphuric acid and potassium dichromate. They were then repeatedly rinsed with deionized water, centrifuged, and the supernatant liquid aspirated, until all acid traces were removed. Because of the high clay content (up to 40% by weight), attempts were made to remove as

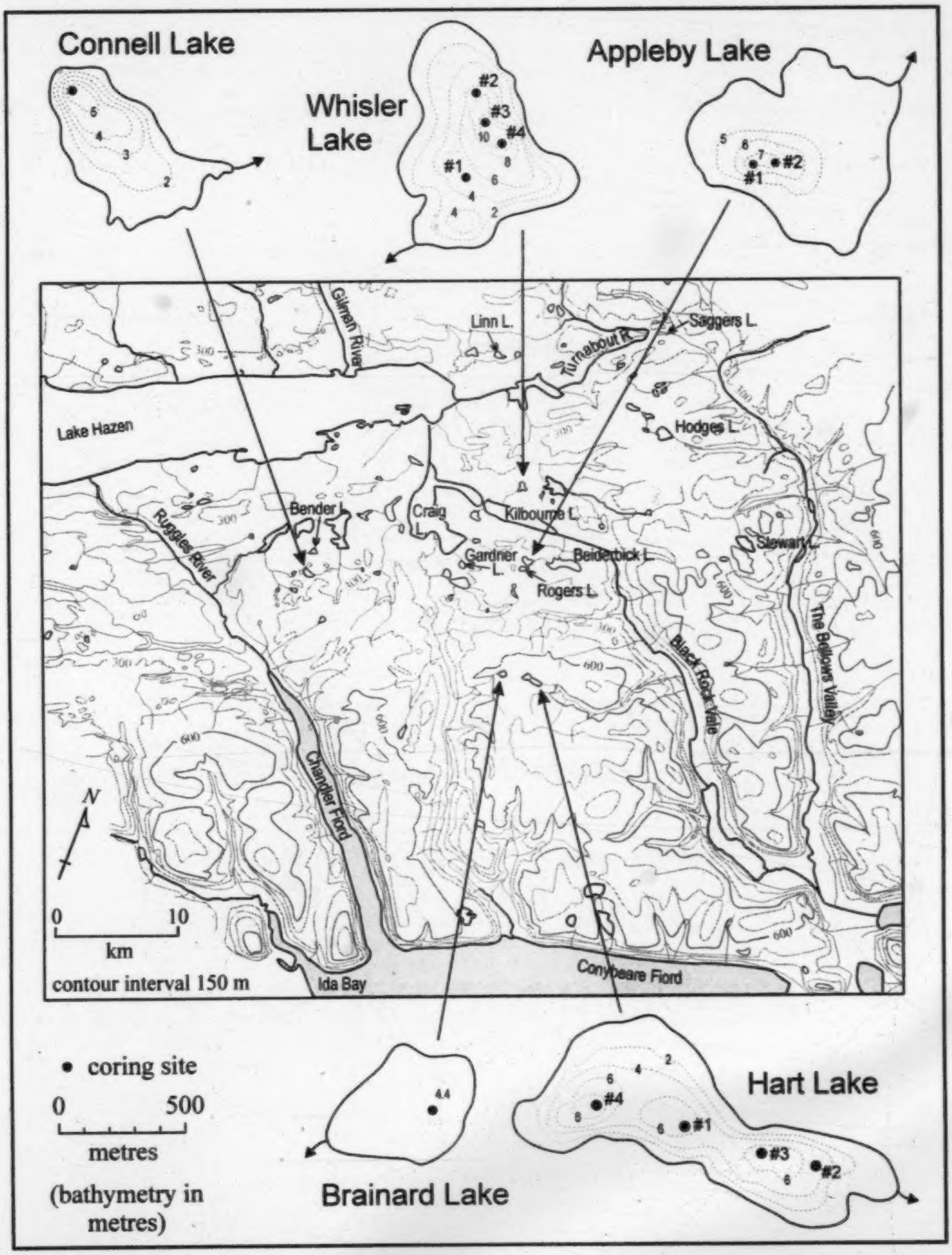

Connell Lake
Whisler Lake
Appleby Lake
#1
#2
#3
#4
Linn L.
Saggers L.
Turnabout R.
Gilman River
Lake Hazen
Hodges L.
Ruggles River
Bender L.
Craig L.
Kilbourne L.
Gardner L.
Beiderbick L.
Rogers L.
Stewart L.
Black Rock Vale
The Bellows Valley
Chandler Fiord
Conybeare Fiord
Ida Bay
300
600
N
0
10
km
contour interval 150 m
• coring site
0
500
metres
(bathymetry in metres)
Brainard Lake
Hart Lake

Figure 2.2 Topographic map of the field area showing location of the lakes cored. Bathymetry and locations of coring sites are highlighted on enlargements of each of the main study lakes.

Table 2.1 Morphometric parameters and chemistry of the study lakes

Lake	Elevation (m asl)	max. Depth (m)	Surface Area (km^2)	Catchment Area (km^2)	pH	Alkalinity (mg/l CaCO$_3$)	dissolved Silica (mg/l)
Appleby	447	7.7	0.488	3.269	7.6	173	>1
Brainard	632	4.4	0.176	2.115	7.7	168	>1
Hart	621	8.5	0.677	4.802	7.5	140	>1
Whisler	245	11.4	0.311	1.466	8.0	187	0.84

much as possible prior to diatom identification. This was achieved by employing a procedure modified from Bates *et al.* (1978), dispersing the samples in a 0.1M sodium pyrophosphate solution and heating them in a water bath, then centrifuging at 2300 rpm for 6 min and aspirating the supernatant (initial runs were also sonicated for 10 minutes to aid dispersion). The procedure was repeated until the supernatant became clear (up to 12 repetitions), after which the sample was rinsed with deionized water, centrifuged and aspirated at least four times. All stages of the procedure were rigorously checked to ensure that diatoms were not being lost.

The remaining sediment was diluted with deionized water, and made up to a known volume (20 ml). The slurry was agitated thoroughly, and then small aliquots were pipetted directly onto circular coverslips. Aliquot volume depended on the diatom concentrations and sediment characteristics of each sample, and ranged from 5 to 100 µl. The sample was immediately dispersed by pipetting deionized water onto the coverslip until it was completely covered, and was then left to evaporate. Once dry, coverslips with a visually uneven distribution of material were discarded, the others were heated at 60 °C for 6 hours, and then mounted on glass slides using DePeX© (refractive index = 1.524).

The highly variable inter- and intra-core diatom abundances (up to six orders of magnitude) prohibited the use of any one counting method. Samples with very high diatom abundances (10^8 - 10^9 VGS), were enumerated by counting absolute diatom numbers in 100 fields of view at x500 power (oil immersion; individual species identification was often made at x1000). Fields of view were selected by starting at the top of the coverslip, moving systematically down the entire coverslip 1 mm at a time, then randomly selecting fields of view across the coverslip. Abundances of individual species were determined using the following equation:

$$N_{\mathit{Total\ species\ X}} =$$

$$\left(n_{\mathit{species\ x\ counted}} * \frac{\mathit{area\ of\ cover\ glass}}{\mathit{area\ of\ field\ of\ view} * \mathit{no.\ of\ field\ of\ views}} * \frac{\mathit{slurry\ volume}}{\mathit{aliquot\ volume}} \right) \div \mathit{sample\ dry\ weight}$$

Total diatom abundances are the sum of all individuals counted. As a check, total diatom abundances from several samples were determined by enumerating an entire coverslip, using very small aliquots (5-10 µl). Differences between the two counting methods, and those determined on recounts of the same slide and on replicate slides, were no more than ± 35%. While this appears large, significant shifts in the diatom records in this study are orders of magnitude, and thus the accuracy of the method is considered acceptable. However, it likely lacks the rigour for inter-study comparisons or in sediments that contain a high species diversity, or relatively invariable diatom abundances (cf. Wolfe 1997). Samples with lower overall diatom abundances ($<10^7$ VGS) were enumerated by continuous scans of a minimum of 1/3 of two entire coverslips. The lower diatom abundances ($0-10^5$ VGS) required absolute counts of 1-5 entire coverslips. Errors associated with these latter two counting schemes were considerably less (±12%).

Species composition was also a concern in designing the counting strategies. In all sediment cores examined, the diatom assemblage was dominated by *Fragilaria* species. Therefore, an attempt was made to count at least 200 non-*Fragilaria* species, so that fluctuations of interesting or ecologically important taxa were not obscured (cf. Battarbee 1986). However, with the exception of Brainard Lake, this was not possible because non-*Fragilaria* species were simply too sparse below the upper few centimetres of sediment.

Species identification was made with reference to Camburn *et al.* (1986), Cleve-Euler (1951-1955), Foged (1972, 1981, 1989), Håkansson & Kling (1989), Kling & Håkansson (1988), Krammer & Lange-Bertalot (1986, 1988, 1991a,b). Chrysophyte stomatocysts were enumerated, but not described.

2.2.3 Radiocarbon Dating

Chronological control was provided by 10 Accelerator Mass Spectrometry (AMS) dates determined on samples of bryophytes (Table 2.2). Species include *Drepanocladus aduncus* var. *kneiffii* (B.S.G.)Mönk., *D. brevifolius* (Lindb.)Warnst., *D. fluitans* (Hedw.)Warnst., *Scorpidium scorpioides* (Hedw.)Limpr. and *S. turgescens* (T.Jens.)Loeske (LaFarge-England, pers. comm. 1995). All are found as submerged or emergent forms around small lakes, ponds or streams (Brassard 1971; Vitt 1975; Janssens 1983; *D. aduncus*

Table 2.2 Radiocarbon date list.

Site	Laboratory Dating No.	Material[a]	Weight used (mg)[b]	Age[c]	Stratigraphy	Core Depth (cm)
Appleby core #2	TO-5146	*Drepanocladus fluitans*	8	1490±50	lacustrine sed.	10
Appleby core #2	TO-5147	*Drepanocladus fluitans*, *Scorpidium scorpioides*	4	1890±50	lacustrine sed.	20
Appleby core #2	TO-5148	*Drepanocladus fluitans*	7	4220±60	lacustrine sed.	70
Appleby core #2	TO-5149	*Scorpidium turgescens*	6	6290±80	lacustrine sed.	102
Appleby core #2	TO-5866	*Scorpidium turgescens*	3	3220±110	diamict	169 - 171
Brainard core #1	TO-5271	*Drepanocladus aduncus* var. *kneiffii*	7	2510±180	lacustrine sed.	14 - 16
Brainard core #1	TO-5272	*Drepanocladus aduncus* var. *kneiffii*	17	3030±110	lacustrine sed.	29 - 31
Brainard core #1	TO-5273	*Drepanocladus aduncus* var. *kneiffii*, *Encalypta* cf. *procerum*, *Orthothecium strictum*	36	2370±60	lacustrine sed.	49 - 51
Brainard core #1	TO-5274	*Scorpidium turgescens*, *Scorpidium scorpioides*	18	7550±80	lacustrine sed.	141 - 144
Connell core #1	TO-5867	*Drepanocladus aduncus* var. *kneiffii*	3	5290±110	sorted sediment	180 - 182
Whisler core#1	TO-4472	*Scorpidium turgescens*		5740±270	lacustrine sed.	27 - 28
Whisler core #1	TO-4473	*Drepanocladus brevifolius*		7600±400	diamict	70.5
Craig Lake Moraine	TO-4206	peat		5970±70	*in situ* peat underlying ice contact delta	

[a] Moss species identified by C. LaFarge-England

[b] Weight of material used to prepare dating target, following pretreatment

[c] Organic samples were subjected to an Acid-Alkali (AAA) pretreatment by IsoTrace Laboratories, designed to remove humic and tannic contaminants and isolate the lignin and cellulose fractions (R.Beukens, pers. comm. 1996)

var. *kneiffii* has previously been unreported from the study area). Discrete mats of bryophytes, up to 5 mm thick, were found throughout many of the lake cores and are considered the product of shore erosion, likely by ice scour or the break-up of anchor ice (bryophyte mats are often seen floating within the peripheral shore lead). Allochthonous redeposition of bryophytes may also have occurred by wind (cf. Glaser 1981) or inwash, although all of the lakes reported here receive only small, ephemeral streams, and most recharge is related to seepage from snowmelt. One sample from Brainard Lake (TO-5273; Table 2) composed predominantly of *D. fluitans*, also contained small amounts of the terrestrial bryophytes *Encalypta* cf. *procerum* Bruch. and *Orthothecium strictum* Lor (LaFarge-England, pers. comm. 1996). Nevertheless, the well-preserved nature of most bryophyte samples (intact leaves attached to stems), suggests that the source is predominantly autochthonous aquatic growth.

Although the bryophytes were likely derived locally, the ubiquitous presence of coal, and the calcareous nature of much of the bedrock and glacial drift, suggests that a reservoir (hard water) effect must be considered in relation to these ^{14}C dates (cf. Abbott & Stafford 1996). It has been suggested, however, that because bryophytes take up CO_2 and not bicarbonate, they are less likely to contain ^{14}C reservoir effects (Mott & Jackson 1982; Glime & Vitt 1984). Studies of bryophytes from Cape Herschel by Blake (1991; Fig. 2.1, site 2), and from northeastern Greenland by Fredskild (1995), support this assertion. However, it has been demonstrated in Alberta that some aquatic bryophytes can contain reservoir-contaminated ^{14}C ages (MacDonald *et al.* 1987, 1991; Aravena *et al.* 1992), and that the ^{13}C composition of these bryophytes cannot be used as the sole criterion to detect reservoir effects.

2.3 RESULTS

2.3.1 Lake Sediments

Within the lake cores, the basal sediments consist of sorted sand and gravel (often displaying current-bedded structures), or a gravel-rich to gravel-poor, massive diamict. The sorted sediments are interpreted to record deposition in extensive proglacial lakes formed during deglaciation. The diamicts, however, are more enigmatic, and a separate, detailed study of their chronology, clast fabrics and biological subfossil remains suggests that they are the product of ice-rafting during deglaciation, although the mechanism by which this occurred is uncertain (refer to Chapter 3).

Above the coarse basal sediments in all the cores is a clayey-silt between 50-150 cm thick. Multiple cores from the same basin reveal similar thicknesses (± 10 cm) for this upper unit, suggesting that sediment focussing is minimal in the absence of major stream inputs. Because of the coring methods employed, it can be assumed that the uppermost sediment was lost. The amount of loss is unknown. Lamoureux & Bradley (1996), using a similar coring method, with the addition of a piston, report the loss of ~50 varves from the top of one core from Lake C2, northern Ellesmere Island (Fig. 2.1, site 3), which corresponds to less than 3 cm of material (Lamoureux, pers. comm. 1996). Logistical and handling problems with the Whisler Lake core resulted in the loss of the upper ~50 cm, while as much as 20 cm may have been lost from the top of the Hart Lake core.

2.3.2 Diatom Stratigraphies

Diatom assemblages from the four study lakes are presented together with a species list and taxonomic authority (including recent nomenclature revisions; Figs. 2.3, 2.4 and 2.5; Appendix 2.1). Certain species and varieties have been grouped together, reflecting uncertainties in identification (i.e., *Achnanthes minutissima* spp.). Others, such as the centric/planktonic taxa have been combined because they are significant as an ecological grouping in the environmental reconstructions, whereas individual species are too sparse to be meaningfully represented. The systematics of the genus *Fragilaria* has been the subject of much research and controversy (Williams & Round 1987, Lange-Bertalot 1989; Round & Williams, 1992); however, within this study I have chosen to retain the *Fragilaria* nomenclature *sensu* Hustedt (1930) and Patrick and Reimer (1966). Distinguishing between the various morphotypes of *F. construens* var. *venter* and *F. pinnata* using the light microscope, was difficult at times. However, as these two taxa commonly occur together (in relatively invariable abundances), and appear to exist under similar ecological conditions, errors in identification should be minimal.

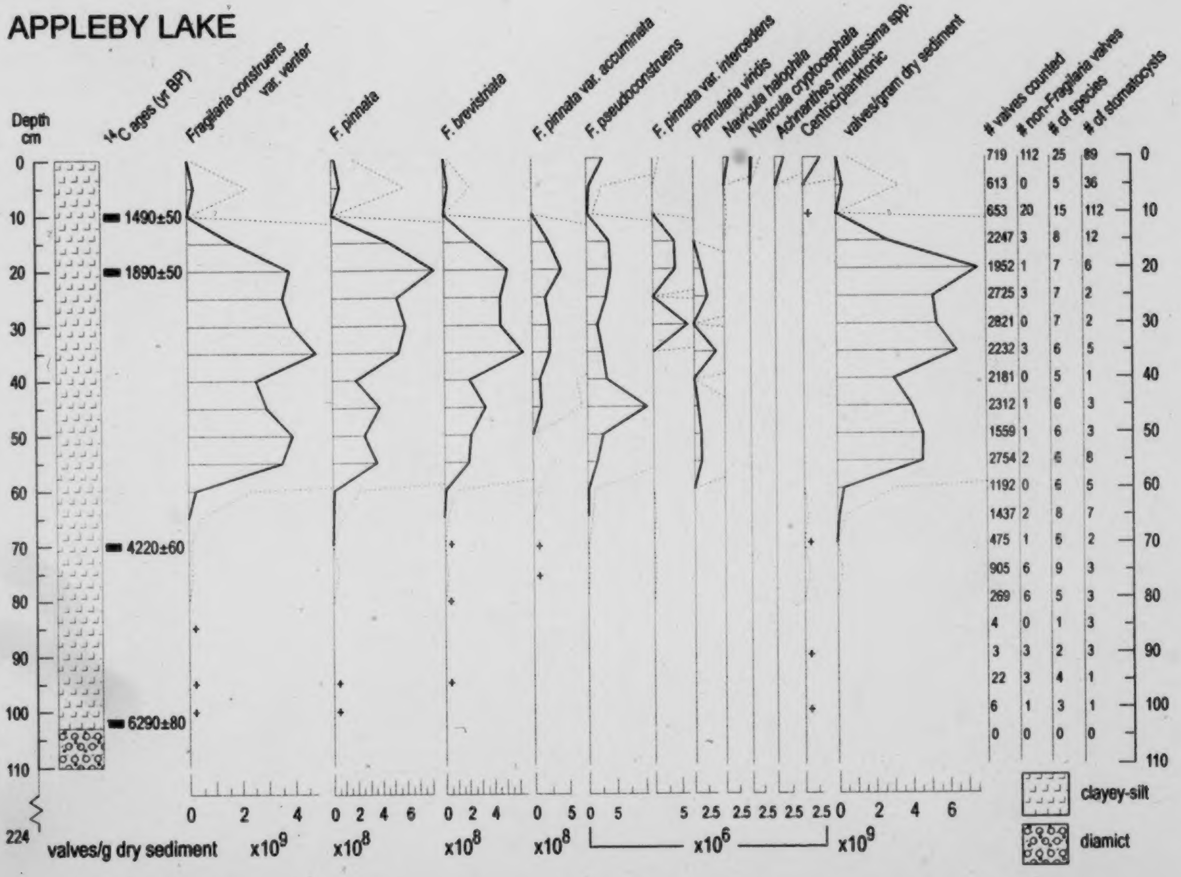


Figure 2.3 Frequency diagram of the diatom assemblage from Appleby Lake core #2. Concentrations are expressed as number of valves per gram dry sediment, listed at the bottom for each species, and arranged from left to right in descending order of abundance. Dashed lines = 10x exaggeration, + signs indicate presence of the species below recordable levels.

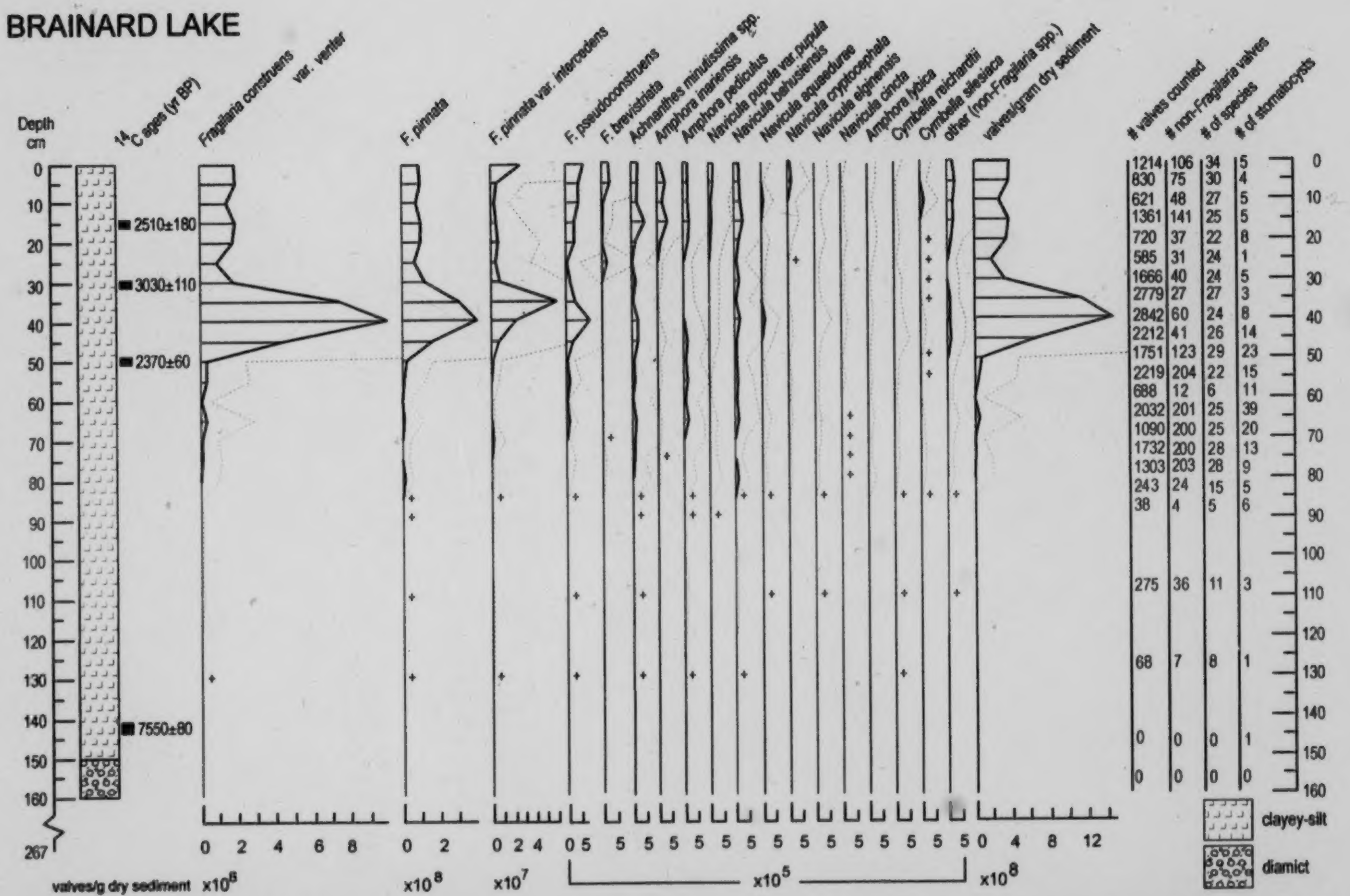
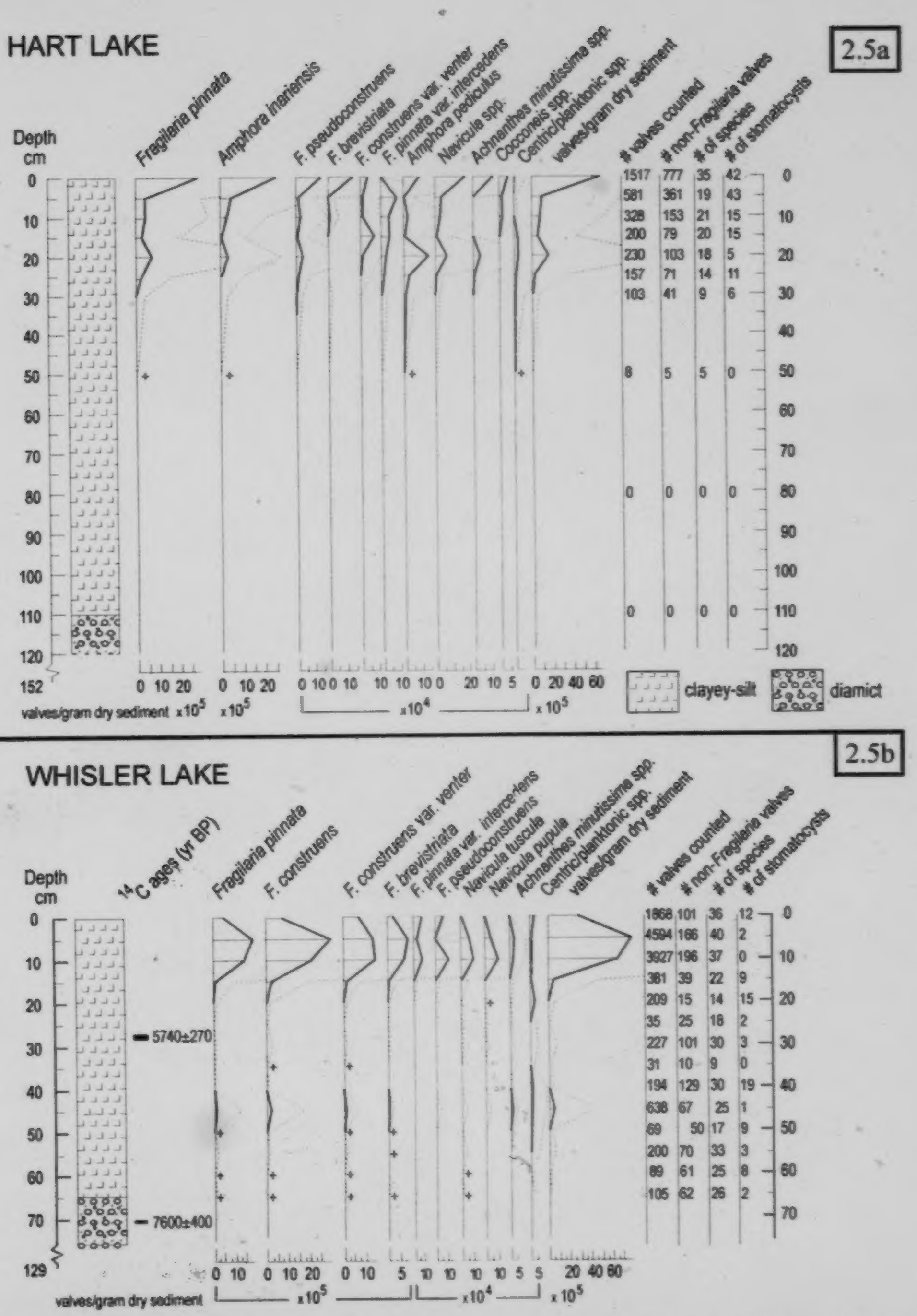


Figure 2.4 Frequency diagram of the diatom assemblage from Brainard Lake core #1. Concentrations are expressed as number of valves per gram dry sediment, listed at the bottom for each species, and arranged from left to right in descending order of abundance. Dashed lines = 10x exaggeration, + signs indicate presence of the species below recordable levels.



# valves counted	# non-Fragilaria valves	# of species	# of stomatocysts
1517	777	35	42
581	361	19	43
328	153	21	15
200	79	20	15
230	103	18	5
157	71	14	11
103	41	9	6
8	5	5	0
0	0	0	0
0	0	0	0

# valves counted	# non-Fragilaria valves	# of species	# of stomatocysts
1868	101	36	12
4594	166	40	2
3927	196	37	0
381	39	22	9
209	15	14	15
35	25	18	2
227	101	30	3
31	10	9	0
194	129	30	19
638	67	25	1
69	50	17	9
200	70	33	3
89	61	25	8
105	62	26	2

Figure 2.5 Frequency diagrams of the diatom assemblages from Hart Lake core #2 (**Fig. 2.5a**) and Whisler Lake core #3 (**Fig. 2.5b**). Concentrations are expressed as number of valves per gram dry sediment, listed at the bottom for each species, and arranged from left to right in descending order of abundance. Dashed lines = 10x exaggeration, + signs indicate presence of the species below recordable levels.

2.3.2.1 *Appleby Lake*

Two cores, 2.37 and 2.24 m in length (core #1 and #2, respectively) were removed from Appleby Lake (Fig. 2.2). The sediments of core #2 (which was used for the diatom analysis) comprised a massive lower gravel-rich diamict (2.24-1.62 m) which graded into a gravel-poor diamict (1.62-1.03 m), the sharp upper boundary of which was overlain by clayey-silt. Five samples of bryophytes were submitted for ^{14}C dating (Table 2.2). The contact between the clayey-silt and diamict is constrained by an age of 6290±80 BP (TO-5149), while a sample from 10 cm depth dated 1490±50 BP (TO-5146; Table 2.2). Sufficient amounts of plant material for ^{14}C dating were very rarely encountered in the diamicts of this and all other lakes cored. The small sample size, and the elimination of the alkali extraction (which generally accounts for the greatest loss of dateable carbon, Beukens, pers. comm. 1996), may account for the anomalously young age of sample TO-5866 (3220±110 BP) taken from the lower diamict, compared to dates on overlying samples (Table 2.2).

Twenty-seven species from 11 genera were identified in the Appleby Lake core (Fig. 2.3, Appendix 2.1). The diatom assemblage is dominated by *Fragilaria* spp. (84% at 0 cm depth; 95-100% at 5-80 cm depth), notably *F. construens* var. *venter*. With the exception of *F. pinnata* var. *intercedens*, there is little in the way of compositional variation with changing diatom concentrations. Species diversity is very low throughout all but the uppermost sediments, as is the number of non-*Fragilaria* species; only *Pinnularia viridis* figures prominently (0.04 - 0.1%; Fig. 2.3). *P. viridis* is a pH-indifferent to circumneutral, epipelic diatom that is widely reported (but rare, generally <2% of any assemblage) from various aquatic habitats throughout Greenland (Foged 1953, 1989) and elsewhere in the Arctic (Hickman 1974, 1975; Douglas & Smol 1993; Brown *et al.* 1994). However, it was not reported by Hamilton *et al.* (1994a) in their survey of several lakes and ponds in northern and central Ellesmere Island.

Total diatom abundance is very low throughout much of the lower 35 cm (65-110 cm depth) of the clayey-silt (7×10^3 - 3.5×10^6 VGS), although there is a slight rise at 75 cm (2.3×10^7). Above this (65-10 cm depth), there is a dramatic rise of diatom concentrations to 10^9 VGS, and then an equally sharp decline to 3.8×10^6 VGS at 10 cm. Concentrations rise somewhat to 3.4×10^8 VGS at 5 cm depth, and then decline to 6.0×10^7 VGS at 0 cm.

Blake *et al.* (1992) report similar peak diatom abundances (up to 6 x 10^9 VGS), and dominance by *F. construens* var. *venter* and *F. pinnata* within the organic section (5-49 cm) of a core from Kap Inglefield Sø, northwest Greenland (Fig. 2.1, site 4). Below this organic section, the sediments were almost devoid of diatoms.

2.3.2.2 *Brainard Lake*

A single core, 2.67 m in length, was removed from near the centre of the lake (Fig. 2.2). The sediments consisted of a massive gravel-rich diamict (2.67-1.95 m) which graded into a gravel-poor diamict (1.95-1.50 m), overlain by clayey-silt. Four samples of bryophytes from the clayey-silt were submitted for ^{14}C dating. The sharp transition from clayey-silt to massive diamict is constrained by a sample from 1.41-1.44 m, dated at 7550±80 BP (TO-5274), while a sample from 0.14-0.16 m yielded an age of 2510±180 BP (TO-5271; Table 2.2). The sample from 49-51 cm (2370±60 BP; TO-5273) is younger than the two dated samples overlying it (Table 2.2, Fig. 2.4), and is thus regarded with suspicion. Reasons for this anomalous date are unknown, as great care was taken to avoid contamination from material dragged along the core barrel. While fragments of two terrestrial bryophytes were included, this sample was predominantly comprised of the aquatic bryophyte *Drepanocladus aduncus* var. *kneiffii*.

Sixty-six species from 13 genera were identified in the Brainard Lake core (Fig. 2.4, Appendix 2.1). Note, the "# non-*Fragilaria* valves" refers only to the number of individuals identified in the initial count (recorded under "# valves counted"; Fig. 2.4). Actual non-*Fragilaria* abundances for samples between 0-80 cm are based on additional scans (and cumulative area), bringing the total number up to 200 non-*Fragilaria* valves. Below 80 cm, and within the sample from 60 cm depth, there were too few non-*Fragilaria* valves to make this counting strategy feasible.

The diatom assemblage is markedly similar to that of the Appleby Lake core: dominated by *Fragilaria* species (87-99%) - particularly *F. construens* v. *venter* and *F. pinnata*, and is compositionally similar throughout large fluctuations in species concentrations. *Fragilaria brevistriata* is, however, considerably less abundant in the Brainard Lake core. Interestingly, there were no centric/planktonic species identified throughout the core. In comparison to

Appleby Lake, the overall diatom concentrations in Brainard Lake are approx. one order of magnitude less, and there is a thinner section of peak diatom abundance (30-50 cm; 10^8 VGS), and a more protracted length of relatively high abundance (0-30 cm; 10^6-10^7 VGS).

2.3.2.3 *Hart Lake*

Four cores, varying in length from 1.52-2.87 m, were removed from sites along the length of Hart Lake (Fig. 2.2). Core #3 was analysed for diatoms, and consisted of a lower gravel-rich diamict (1.52-1.10 m), directly overlain by a clayey-silt. No samples from the cores of this lake were submitted for ^{14}C dating.

Forty-three species from 19 genera were identified in the Hart Lake core (Fig. 2.5a, Appendix 2.1). The diatom assemblage is dominated by *Fragilaria pinnata* (34-49%) and *Amphora inariensis* (20-52%; a form often reported from oligotrophic, nordic alpine waters). Differences in species composition between Hart and Brainard lakes likely reflect the greater size and depth of Hart Lake, and consequently, a more persistent ice floe cover. Unlike Brainard Lake, Hart Lake contains centric diatom species (predominantly the tychoplanktonic *Aulocoseira* spp., and also some *Cyclotella* spp.). Between 46 and 53 cm core depth, there are several mats of bryophytes. With the expectation of there being epiphytic diatoms associated with the bryophyte mats, three samples were analyzed, however, no such assemblage was observed (results are presented only for the middle sample, 50 cm; Fig 2.5a). Sediments with discrete mats of bryophytes from the cores of other lakes similarly displayed no distinct diatom assemblage or increased abundance. Approximately 20 cm of loose unconsolidated sediment was discarded from the top of core #3 upon splitting. Comparisons of the diatom concentrations in other Hart Lake cores (and the adjacent Brainard Lake core) suggest that at least this much upper sediment was lost. The trend in diatom concentrations appears similar to that seen between 85-45 cm in the Brainard Lake core.

Three unique diatoms were found within the Hart Lake core. Based on comparisons with Mölder & Tynni (1968) and Tynni (1982), these are interpreted to be the marine forms, *Actinoptychus undulatus* (2 valves) and *Distephanus speculum* (1 valve), possibly of Tertiary (or greater) age. The marine nature of these diatoms is incompatible with the fresh water, intermontane origin of Tertiary materials in this region (Miall 1979). While redeposition from other strata cannot be ruled out, it does support arguments made in Chapter 4 of this thesis, that much of the bedrock exposed around the Hart Lake basin, consists of late-Mesozoic marine deposits.

2.3.2.4 *Whisler Lake*

Four cores, varying in length from 2.85-1.29 m, were removed from Whisler Lake (Fig. 2.2). Core #1 was analysed for diatoms, and was composed of a lower pebble-rich diamict and open to closed-work gravel deposit (1.29-0.65 m) directly overlain by a clayey-silt. Three samples were submitted for ^{14}C dating, but the uppermost sample (0.06 m) contained insufficient carbon for analysis following pretreatment. The other two dates, TO-4472, 4473 (Table 2.2) both have large standard errors, likely reflecting the small amount of material submitted for dating (a single branch of moss comprised each sample). The date of 7600±400 BP (TO-4473) provides a maximum age for the diamict/clayey-silt transition.

Ninety-three species from 25 genera were identified in core #1 (Fig. 2.5b, Appendix 2.1). While this lake has the greatest species diversity of all four lakes, its assemblage is dominated by the fewest number of species (four, all *Fragilaria*; 80-96% between 0-20 cm, 28-90% between 25-65 cm). Only three non-*Fragilaria* species occur above trace concentrations throughout the core: *Navicula tuscula* (0.8-1.6%), *N. pupula* (0.4-1.9%) and *Achnanthes minutissima* (0.2-0.6%). Other species occur in concentrations of 10^4 - 10^5 VGS, but quite sporadically, or in only one sample. *N. tuscula* was not found in any of the other lake cores investigated, but is reported at low abundances (0.25-2.56%) by Hamilton *et al.* (1994a; as *Aneumostus tusculus*) in 4 of 21 lakes and ponds sampled. It is also reported as a predominantly epilithic and epipelic form in ponds at Cape Herschel (Douglas & Smol 1994), as a common, but rare, subfossil and modern alkaliphilous form in several Greenland lakes and ponds (Foged 1953, 1955, 1989), and as very common in the surface sediments of a small (1.25 km^2) alkaline lake in northern Alaska (pH 8.0, max. depth 15.5 m; Foged 1971). *N. pupula* is found at moderate abundances (10^5 VGS) throughout much of the Brainard Lake core (Fig. 2.4), at low abundances (10^3-10^4 VGS) throughout the Hart Lake core, and only within the surface sample of Appleby Lake (1.3 x 10^5 VGS). It was not reported from Cape Herschel, but is considered a rare, pH indifferent form in lakes and ponds on Greenland (Foged 1953, 1955), and occurs commonly (but in low numbers) in a survey of the epipelon of lakes from the southern Yukon (Hickman 1975). Hamilton *et al.* (1994a) report its uncommon occurrence and low abundance (<0.7 %) from 2 ponds and a small lake south of Craig Lake (Fig. 2.2). *A. minutissima* is widely reported from all manner of Arctic lakes, ponds and streams and is often most prominent in the littoral communities (Foged 1989; Douglas & Smol 1993; Hamilton *et al.* 1994a; Ludlum *et al.* 1996).

2.4 DISCUSSION

2.4.1 Nature of the Diatom Records.

Few studies have examined the diatom-based sedimentary record from lakes in the Canadian High Arctic. This study is notable because it examines small, weakly alkaline basins without major stream inputs. The diatom records share both similarities and marked differences with studies conducted on neutral to acidic basins elsewhere in Arctic Canada (i.e. Smol 1983; Wolfe 1994), and on sites from around Greenland. They are most broadly similar to results reported from Kap Inglefield Sø, a small, shallow (5.6 m), weakly alkaline (pH 7.4) basin (Blake *et al.* 1992; Fig. 2.1, site 4).

One hundred and forty-four species from 28 genera are reported from the cores of the four lakes. Of these, 8 % occur in all four basins, while 64% occur only in any one of the four basins (Appendix 2.1). The diatom assemblages are dominated by benthic alkaliphils, notably those of the genus *Fragilaria*. Other forms encountered are predominantly circumneutral or indifferent. A benthic *Fragilaria* assemblage has been widely reported as dominating the epipelon from early postglacial sediments in temperate and alpine regions (i.e. Haworth 1976; Whitehead *et al.* 1986; Hickman & Schweger 1991), and from sites in the Arctic on both carbonate and acidic terrains (Foged 1953, 1955; Smol 1983; Lemmen *et al.* 1988; Blake *et al.* 1992). The latter is interpreted to reflect initial leaching of base cations following deglaciation (cf. Pennington *et al.* 1972; Likens & Davis 1975). While this assemblage is generally held to be an early colonizing group, its predominance throughout the Arctic sediment records likely reflects the severity of the environment, notably the restrictions on photosynthetic activity imposed by the short summer melt season and its associated lake ice and snow cover (Smol 1988). Observations from Char Lake, Cornwallis Island (Fig. 2.1, site 5) show that only 34% of the potential annual surface light available under ice-free conditions penetrates to the water (Schindler *et al.* 1974; Welch *et al.* 1987). This amount can be further affected by the thickness of both snow and ice cover, as well as by the structure of the surface ice floe (i.e. white ice compared to black ice). Welch *et al.* (1987) predicted that 43% of incident light will pass through a 1 m thick floe of candled black ice. In this study, the lakes occupy an altitudinal transect with a significant climatic gradient (where the lower lakes generally clear of ice earlier). Hence, it could be argued that those species more abundant in lakes at higher altitudes (*F. construens* var. *venter* and *F. pinnata*) are better adapted to low light conditions. The connection is unlikely to be quite so direct, as other factors such as depth and opacity of the water column are likely to be equally limiting. Temperature variations within Arctic lakes are small, and thus unlikely to be a significant constraint on resident benthic diatom populations (Smol 1988). The lakes studied are also chemically similar (Table 2.1). Differences in the relative composition of the various *Fragilaria* species within the four lakes are thus considered to record unknown environmental gradients. Asymmetric/misshapen valves (predominantly *Fragilaria*) were found throughout all cores (<1 % of all valves recorded). While silica concentrations may limit overall productivity (Hickman & Reasoner 1994), or cause changes in valve dimensions, it is unlikely to be the cause of the raphe, striae and asymmetric valve abnormalities seen here. This is especially true considering the presence of the heavily silicified *Aulocoseira* valves.

While there appears to be little in the way of overall diatom succession, there are several noted changes in the lake assemblages. For example, in Appleby Lake, three species (*F. pinnata* var. *accuminata*, *F. pinnata* var. *intercedens* and *Pinnularia viridis*) occur only during periods of greater diatom abundance (Fig. 2.3), and therefore, likely record increased light conditions. The rise of these three forms does not appear to affect the distribution of other dominant taxa, and thus competitive stresses are unlikely. Competition cannot be altogether ruled out in all of the lake records, and may account for the general decline in some species (notably *Amphora inariensis*, *A. pediculus*, *Navicula cryptocephala* and *Cymbella silesiaca*) within Brainard Lake during peak diatom abundance (Fig. 2.4). Other factors, such as the shallower water depth and hence, potentially increased light availability, may explain the more abundant and consistent occurrence of *F. pinnata* var. *intercedens* within Brainard Lake, during periods of high to moderate productivity.

2.4.2 Ice-cover Proxy Record

Based on studies of the diatom assemblages from Rock Basin Lake, east-central Ellesmere Island (Fig. 2.1, site 6), Smol (1988) developed the hypothesis that fluctuations in both diatom productivity and composition could be related to the extent of the surface ice cover during the summer. Simply put, in warmer years when the ice floe melts completely, it would be expected that benthic regions would record higher productivity and increases in the more deep water and planktonic taxa. In colder years, when only a peripheral shore lead formed, productivity would be low, and aerophilic and shallow water taxa would dominate the benthic thanatocoenosis (subfossil assemblage). Based on this model, the dramatic shifts in diatom abundance recorded in this study (Fig. 2.6), would reflect changes in productivity that are controlled by the extent of the summer ice-cover. Therefore, high diatom abundance would correlate with "warm" summers, whereas a low abundance would correlate with "cold" summers. While it is discussed in terms of temperature changes, it should be recognized that the diatoms serve only as a proxy-ice cover record, and that changes in lake ice cover may be related as much to the amount and seasonal variations in precipitation (i.e., snow vs. rain), as they are to temperature, or local topoclimatic differences.

The difference between the results of this study and those of Smol (1983, 1988), is the rarity of littoral, shallow-water forms throughout the Holocene, regardless of hypothetical changes in ice cover. This is despite the fact that discrete habitat sampling (rock scrapes, aquatic plants, and surface sediments) around the margins of these and other lakes and ponds throughout the region reveal a high productivity and diversity (commonly 40 - >130 taxa) most of which are absent from the benthic records (Smith, unpublished data; Hamilton *et al.* 1994a). Nor does there appear to be any relation between intervals of assumed reduced ice cover (warm) and increases in planktonic taxa (note in particular Appleby Lake, Fig. 2.3). In fact, it is misleading that on a percentage basis, planktonic taxa are most prevalent during periods of low productivity - reflecting the absence of an *in situ* benthic population, and the

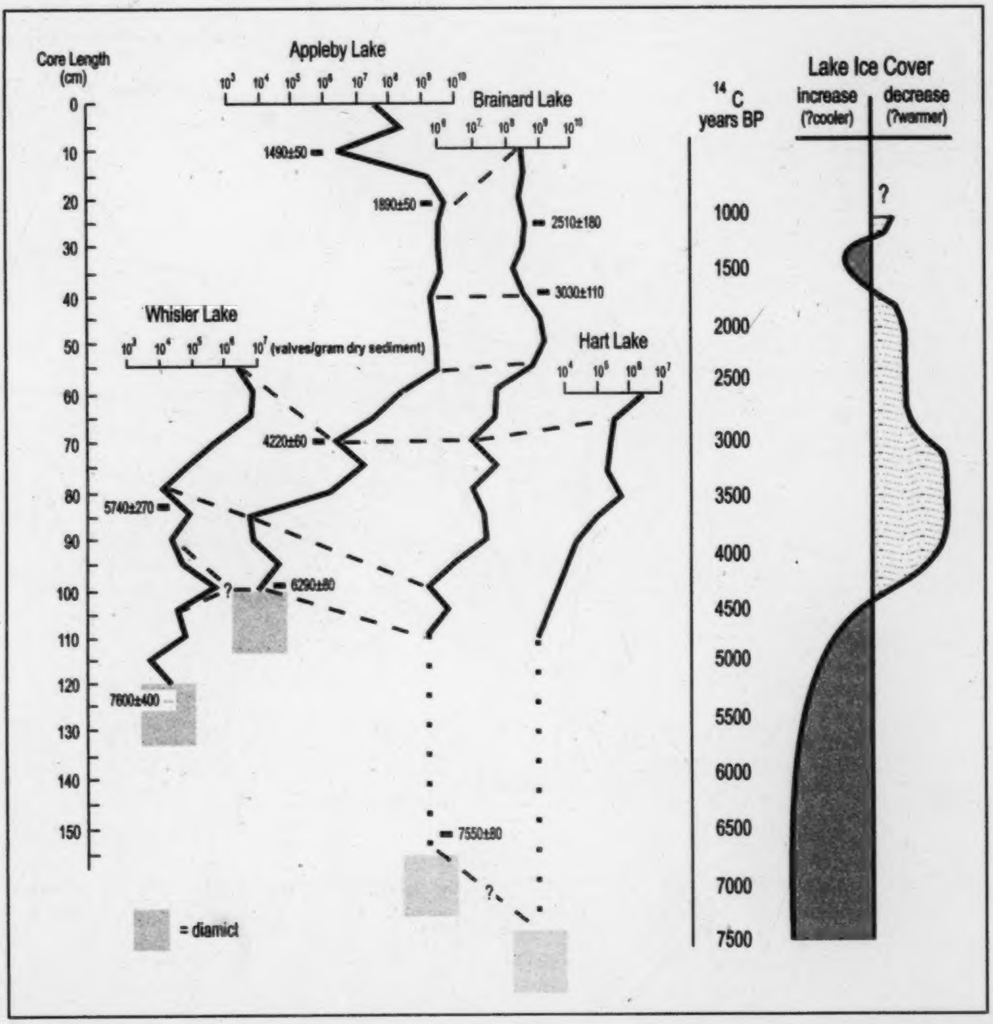


Figure 2.6 A provisional paleoenvironmental reconstruction. Diatom abundances (valves per gram dry sediment) from the four study lakes are plotted and chronostratigraphically linked. Variations in diatom abundance are related to changes in summer lake ice cover. This yields a provisional paleoenvironmental reconstruction (graphically represented at right) in which deglacial, cooler conditions persisted until ~4 ka BP, after which there was a regional warming between 4 and 3 ka BP. Between 3 and ~1.8 ka BP, conditions at lower elevations remained warm, while those around lake basins at higher elevations (i.e, Brainard Lake) appear to have cooled, interpreted to reflect a lowering of the snowline. Following ~1.8 ka BP, conditions at lower elevations also cooled, but then began warming after 1.5 ka BP.

likely redeposition of planktonic forms from the shore lead and possible sub-ice floe growth (Spaulding *et al.* 1993; Wolfe 1996b). Similarly, the stomatocyst remains of chrysophytes, a predominantly planktonic alga, are found in low concentrations at all four lakes, and generally occur in greatest abundance during interpreted "cold" periods. Their relative abundance during these periods may reflect their adaptive resting spore (stomatocyst), motility, and nutritional strategies. The concentrations of stomatocysts observed are comparable to those reported in Kap Inglefield Sø (Blake *et al.* 1992), but are considerably lower than those observed in lakes and ponds around Cape Herschel (Smol 1983; Duff & Smol 1988; Duff *et al.* 1992) and other Arctic lakes and ponds (Hobbie 1973; Wolfe 1994).

Although the rarity of littoral and planktonic forms in the benthic records may be due to dissolution and/or differential preservation, this is considered unlikely because thin spines and fine ornamentation were observed on stomatocysts throughout each of the four cores. Similarly, degraded and broken valves were extremely rare (<1 %). Herbivory or grazing pressures by testate amoebae (genus *Paraquadrula* and *Trinema* observed in some littoral samples), ostracodes (abundant remains found throughout the Brainard and Appleby cores - undetermined in the Whisler and Hart Lake cores), and other invertebrates such as *Daphnia* may also play a role. Cryoturbation and other periglacial processes operating within the margins of the lakes may also play a role (non-sorted stripes and circles observed down to depths of 2 m). Ultimately, the absence of littoral and planktonic taxa may reflect the lack of effective transport mechanisms operating within these lakes. There are only small ephemeral streams, fed by snowmelt, entering these lakes. Inflowing meltwater and lake ice melt have been shown to form a cohesive unit that travels through small lakes such as these as a mostly intact sub-ice flow (Bergmann & Welch 1985). Flow rates are sufficient to carry much of the finer (diatom-sized) material in the sub-ice flow, which may become flushed from the basin (cf. McKnight *et al.* 1990). Wind-driven turbulence can contribute to littoral diatom redistribution, however, the study area is noted for its lack of strong winds (Thompson 1994). Nor would these lakes experience the katabatic winds that Rock Basin Lake likely would. The persistence of extensive ice floes throughout much of the melt season would also dampen a wind effect. Wind, however, may be responsible for the occurrence of the various tychoplanktonic *Aulocoseira* taxa observed in this study. There is a discontinuous, though often notable, circulation around the perimeter of lakes within the shore lead, and this may create enough turbulence to keep these species in suspension.

2.4.3 Provisional Paleoenvironmental Reconstructions

Assuming that the changes in diatom productivity noted in this study can be linked to changes in surface ice cover, and are thus a proxy environmental record, the following reconstruction is proposed (Fig. 2.6).

Mapping of the regional glacial geomorphology, and dating of raised ice-contact, marine deltas, indicates that around 8 ka BP glaciers occupied most of Lake Hazen Basin and inner Hazen Plateau, extending to the mouth of the Ruggles River and Black Rock Vale (England 1978, 1983; refer to fourth paper of this thesis). Much of the outer Hazen Plateau surface was likely covered by plateau ice caps. Retreat from this position (which is not considered to mark the limit of ice during the LGM) occurred slowly. This is indicated by a date of 5970±70 BP (TO-4206) on a peat deposit overlain by an ice-contact delta between Craig and Kilbourne Lake, and a date of 5290±110 BP (TO-5867, Table 2.2) from sorted sediment in Connell Lake (interpreted to have been deposited in a proglacial lake). Subsequent glacial retreat occurred rapidly as shown by a date of 4980±70 BP (GSC-3451) on a peat deposit, 1.5 km northwest of Hazen Camp (Fig. 2.1, site 1), indicating ice no longer occupied Lake Hazen. Retreat history of the plateau ice caps is poorly constrained, but they are considered to have persisted through the mid-Holocene, likely due in part to their favoured topoclimatic position (cf. Bradley & Serreze 1987). The oldest dates on lake sediments from this study are 7550±80 and 6290±80 BP (Table 2.2), from sediments which immediately overly the diamicts in Brainard and Appleby Lakes, respectively. These dates provide a minimum estimate for deglaciation which, dependent upon the interpretation of the diamicts (i.e. till and/or ice-rafted), may have occurred earlier. Interpretation of the 7.6 ka BP date in Whisler Lake is uncertain, as the bryophyte has clearly been reworked. Therefore, 7.6 ka BP is a minimum estimate for deglaciation, but only a maximum age for the contact between the diamict and overlying clayey-silt. During deglaciation, the climate of the region would certainly have differed from that of today. Glaciers infilling Lake Hazen Basin and lower snowlines on Hazen Plateau would have reflected much of the incident radiation (cf. Koerner

1980; Bradley & Serreze 1987), which today is concentrated along the NW facing slopes of the plateau. The consequence of this would be that the winter temperature inversion existing within the basin today would have persisted into the summer periods during deglaciation. Under these conditions, lakes are likely to have retained a near-permanent ice cover, accounting for the relative absence of diatoms in the study lakes in the early to mid-Holocene.

The first marked rise in diatom numbers occurs in Whisler Lake between 40-50 cm, prior to 5.7 ka BP (Fig. 2.6). This peak is not recognized in Appleby Lake which has sediments bracketed by dates of 4.2 and 6.3 ka BP, although this may reflect problems of dating and sampling resolution. Within Brainard Lake, diatom concentrations rise slightly from 4.09×10^5 to 1.49×10^6 VGS between 90 and 110 cm (Fig. 2.6). Just as there are striking topoclimatic variations in the region today, similar conditions in the past may have allowed for the local "warming" around Whisler Lake, while other lakes retained a more continuous ice cover. The persistence of plateau ice caps would also have created locally cooler conditions around Brainard and Hart lakes (cf. Bradley & Serreze 1987).

The rise in diatom concentrations in Appleby, Brainard, Hart and Whisler Lakes at 80, 85, 30 and 20 cm depth, respectively, appears to herald a climatic amelioration from previously "cold" conditions (marked by the very low or even trace diatom occurrences; Fig. 2.6). The date of 4220±60 BP from 70 cm depth in the Appleby Lake core provides the most direct control on this period (~5-4 ka BP?), and suggests that it is approximately coeval with the final retreat of glaciers from Lake Hazen (4980±70 BP, GSC-3451; Blake 1985). LaFarge-England *et al.* (1991) who studied an extensive peat deposit around Piper Pass, ~35 km northeast of Lake Hazen (Fig. 2.1, site 7) indicate ample meltwater availability from 6.4-3.4 ka BP. The seeming delay in the onset of warming (increased melt) between their site and Appleby Lake, may reflect localized moderating influences of glaciers.

Shortly after 4.2 ka BP, diatom concentrations rise abruptly in Appleby Lake (2-3 orders of magnitude; Fig. 2.6). Within Brainard Lake, a similarly dramatic rise in diatom abundance occurs at 50 cm. The date of 2370±60 BP (Fig. 2.4, Table 2.2) is rejected as likely being in error, and instead this rise is considered roughly coeval with that in Appleby Lake. This peak diatom abundance within Brainard Lake is sustained until just prior to

3030±110 BP, when concentrations drop by one order of magnitude (Fig. 2.6). This same trend is not seen in the Appleby Lake core (Figs. 2.3 and 2.6), where high productivity is maintained until ~1.9 ka BP. Differences between the two lake records likely highlight the topoclimatic controls within the region, reflecting Brainard Lake's higher elevation and location on Hazen Plateau (Fig. 2.7). The period ca. 4-3 ka BP is thus considered to mark the warmest conditions throughout the Lake Hazen and Hazen Plateau regions. On average today, Lake Hazen clears of ice only once in 20 years (Wissink, pers. comm. 1994). During this proposed warming period (~4-3 ka BP), Lake Hazen may have cleared more often, allowing a greater warming of the surface waters, which in turn may have further augmented the regional warming, by ameliorating the early fall climate, analogous to the large-lake system on southern Baffin Island today (Jacobs & Grondin 1988).

Within Appleby Lake, productivity was maintained at a high level from ~4 ka BP until shortly after 1890±50 BP. There is evidence to suggest that during this period, productivity may actually have increased in the latter stages, notably the rise in overall diatom abundance, and specifically that of *F. pinnata* var. *accuminata*, *F. pinnata* var. *intercedens* and *Pinnularia viridis* (Figs. 2.3 and 2.6). Differences in sedimentation rates, however, may be significant here in their effect on calculated diatom abundances, but do not explain the unique occurrence of these three species. Within Brainard Lake, diatom abundances appear more or less constant, with little change in overall species composition, from 3 ka BP onwards. The top of the Brainard Lake core is undated, but is considered to date ~1.9 ka BP based on comparisons with the Appleby Lake record, notably the absence of a similar marked decline in diatom abundance. The changes noted are considered to record continued warm conditions within the Lake Hazen Basin from ca. 4-1.9 ka BP, while on Hazen Plateau, conditions cooled somewhat from their peak (4-3 ka BP), yet still maintained a relatively high productivity and diversity of forms. This may record a general lowering of the regional snowline, and a decreased frequency of open-water conditions on Brainard Lake. Conditions may have been analogous to those experienced during the summer of 1992, when Appleby Lake cleared of ice by late-July, yet the snowline never rose above 650 m on the plateau, and hence Brainard Lake may only have formed a small peripheral shore lead (i.e., Fig. 2.7).

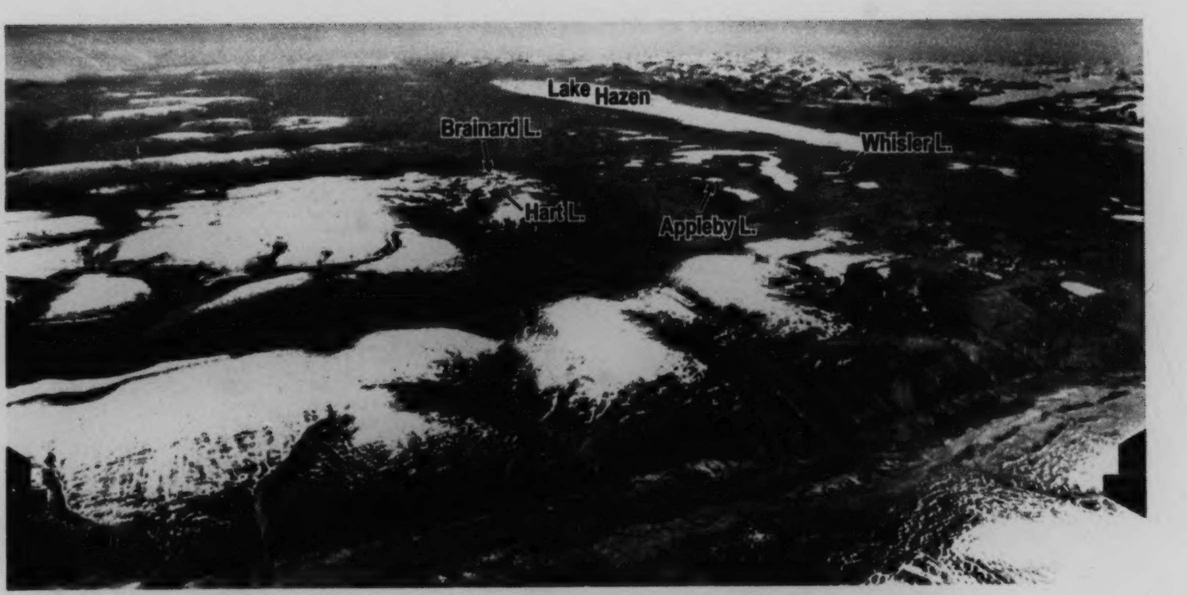


Figure 2.7 Oblique air photograph (taken June 24, 1950) looking west across Hazen Plateau and Lake Hazen Basin towards the Grant Land Mountains, demonstrating the local topoclimatic variation. Lake basins at lower elevations are clear of snow, but retain >95% ice cover, and only a thin shore lead. At the same time, Hazen Plateau remains extensively snow covered, with complete ice cover on the lakes. Breakup of ice floes on lakes at lower elevations is subsequently rapid, while it may be delayed 2-4 weeks on Hazen Plateau, or persist throughout colder summers. [Air photograph T397R-133 © (1950) Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada]

The late mid-Holocene warm period proposed here, also coincides with the findings of regional archaeological surveys that indicate peoples of the Independence I culture (4.5-3 ka BP) made extensive use of the region surrounding Lake Hazen during this time (Sutherland 1991a, Dick *et al.* 1994). Occupation of the Lake Hazen region also appears to have persisted well after it had begun to decline in coastal areas of western Ellesmere Island and northern Greenland (~3.5 ka BP; Sutherland 1991b; Dick *et al.* 1994), suggesting a disparity in the coastal and interior climates.

Shortly after 1.9 ka BP, diatom abundances within Appleby Lake decline sharply, bottoming out at 10 cm depth, dated 1490±50 BP, then maintain lower levels through the upper sedimentary record (Fig. 2.6). The severity of the conditions at this time is reflected in the returned dominance of a few benthic *Fragilaria* forms, and the rise in stomatocysts (Fig. 2.3). Archaeological records (Late Dorset) from the region are sparse from ~2.2-0.9 ka BP, supporting the notion of a climatic cooling (Sutherland 1991a,b; Dick *et al.* 1994).

Although there is a rise in the "# of non-Fragilaria species" at 10 cm, there were only single valves of each species identified, and the possibility must be entertained that these were redeposited and/or reworked from shallower sediments. Other Arctic lake and pond sediment studies have identified a dramatic increase in diatom species diversity in the uppermost sediment, interpreted to reflect recent, possibly anthropogenic, warming (Douglas *et al.* 1994; Doubleday *et al.* 1995). However, this hypothesis cannot be tested here because the coring techniques employed did not preserve the uppermost sediment in an undisturbed state.

2.4.4 Comparisons with other Regional Paleoenvironmental Records

The unique topoclimatic setting and continentality of the Lake Hazen region may be responsible for much of the seeming incongruity between the paleoenvironmental reconstructions proposed here and those reported from other sites and records in the High Arctic (cf. Bradley 1990). The disparity between the records may also in part reflect the bias of previous climatic reconstructions from coastal areas, and the effects of local marine conditions upon them. Different proxy-environmental records are subsequently contrasted with the lake story developed here, and possible explanations for the differences proposed.

2.4.4.1 *Early to mid-Holocene Records*

Smol's (1983) interpretation of the diatom stratigraphy from Rock Basin Lake suggests a climatic warming from ~7.5 to 4 ka BP, followed by an abrupt climatic deterioration that has persisted since. Pollen analysis of the same core by Hyvärinen (1985), identified four pollen zones that are primarily successional, not climatic, in nature. Only the uppermost pollen zone (zone 4: ~3500 BP to present) may be climatically significant, indicating the end of vegetational stability and spread of drier and poorer heath-type vegetation (Hyvärinen 1985).

The interpretation of pollen, diatom and invertebrate microfossil remains in the Kap Inglefield Sø core, suggests warm conditions from ~6.5 to 4 ka BP, followed by a very pronounced cooling, in which the lake is considered to have retained a near continuous ice cover (Blake *et al.* 1992). However, the dating control on the core must be treated cautiously, as ages were determined on bulk sediments, and one sample was extracted from sediments containing sand and pebbles, suggesting the possible reworking of littoral sediments. Errors associated with bulk sediment dating in Arctic lakes are now legion (cf. Retelle *et al.* 1989; Short *et al.* 1994; Lamoureux & Bradley 1996), reflecting low organic carbon contents, contamination by ^{14}C-depleted terrestrial organic material, and the dissolution of inert carbonate materials.

Studies on the diatom assemblages from ponds on Cape Herschel (Douglas *et al.* 1994) yielded two striking conclusions: 1) that there has been a marked post-18th Century warming allowing the proliferation of new taxa, and 2) that this warming is unprecedented as indicated by the absence of any significant change in the diatom assemblages or abundance back to 8 ka BP (oldest basal date from the three ponds). Several independent studies have confirmed the former conclusion (cf. Overpeck *et al.* 1997), while reconciling the latter conclusion with other regional paleoenvironmental records is difficult.

Further evidence of mid-Holocene warm conditions is provided by a date of 5730±70 BP on algae from within the lake ice (4 m depth) of a 5.45 m deep lake in central Ellesmere Island (Fig. 1, site 8; Blake 1989a). This lake presently remains frozen to its bed year round, but presumably experienced some degree of melting in the mid-Holocene (Blake 1989a)

Resolving the disparities between the lake records from Cape Herschel and Kap Inglefield Sø with those from the Lake Hazen region, likely highlights differences in the local climatic and topoclimatic regimes, and the influence of the North Open Water (NOW) polynya (Fig. 2.1). NOW is a recurrent polynya, or area of low ice concentration and thin ice cover, which forms during the winter and spring (Müller *et al.* 1980). Because of its relatively warm temperatures (up to 20°C compared to adjacent areas of fast ice), and the presence of open-water, it provides an abundant moisture source for the immediately surrounding regions (Müller *et al.* 1980; see also Hjort 1997). Increased late-winter and spring snowfall would increase the albedo, delay summer snowmelt, and allow lake ice cover to persist longer, if not through, the entire summer melt season. NOW is also responsible, particularly in the summer, for abundant and persistent fog development along the adjoining coastal areas. The presence of persistent fog along coastal areas such as Cape Herschel and Kap Inglefield would greatly diminish the potential summer radiation, creating cooler conditions, in which surface ice covers on lakes and ponds would be expected to persist longer through the summer. Comparisons between coastal and interior regions adjacent the Northeast Water polynya, situated northeast of Greenland, demonstrate the regional disparity in radiation balance; July-August temperatures of 1.1 - 2.3°C and degree-days above 0°C of 213-222 for two coastal sites, compared to 4.2°C and 305 degree-days for an interior site (Bay 1992). The same sharp climatic differences exist today between Cape Herschel and sites immediately northwest of this (cf. Edlund & Alt 1989). Abundant archaeological records around Buchanan Bay (Fig. 1, site 9), and an absence of similar sites around Cape Herschel, also argue that such climatic disparities existed in the past (Maxwell 1985; Schledermann 1990)

Historical changes in the extent and influence of the NOW are uncertain. Inferences, however, can be drawn from proxy records. Between 5 and 3.5 ka BP, the least negative $\delta^{18}O$ values for the entire Holocene were recorded in the Devon Ice Cap (Koerner 1989). These are interpreted to show maximum winter and/or mean annual warmth, and the corresponding conductivity maximum reflects increased deposition of marine salts from seasonally longer periods of open water in Baffin Bay, specifically the NOW polynya (Koerner 1980). Dyke *et al.* (1996a) also suggest that a mid-Holocene rise in the number of bowhead whale (*Balaena mysticetus*) fossils around Jones Sound may indicate an expansion of the NOW. Other evidence supporting the predominance of the NOW polynya during the mid-Holocene are the presence of boreal-subarctic and diverse arctic mollusk assemblages extending northwards to Cape Herschel (Dyke *et al.* 1996b), and along western Greenland, indicating local coastal waters warmer than present (Funder & Weidick 1991).

2.4.4.2 *Post 3 ka BP and the Neoglacial*

A sharp decline in diatom abundance occurred ~3 ka BP in Brainard Lake, but not in Appleby Lake (Fig. 2.6). This is attributed to a lowering of the snowline. Such a condition could result from an increase in total snowfall and/or a decrease in summer temperatures. There is abundant evidence from Ellesmere Island and western Greenland that glaciers underwent considerable advances post 3 ka BP from ice margins well behind those of today (>18 km; Blake 1989b; Weidick *et al.* 1990; van Tatenhove *et al.* 1996). How much of this is attributable to changes in precipitation is uncertain as the ice core records indicate no clear trend during this period (Paterson & Waddington 1984; Meese *et al.* 1994; Fisher *et al.* 1995). Evidence of increased polar cooling between 2400-3100 calendar years BP is suggested by O'Brien *et al.* (1995) based on sea salt and terrestrial dust records of the GISP2 core. Also, percent-melt records from the Agassiz ice cap show a decline in summer temperatures since ~5 ka BP, especially after 2 ka BP (Koerner & Fisher 1990). The actual amount of cooling is difficult to quantify (suggested as -2°C through the Holocene, Koerner & Fisher 1990), as changes in sea ice extent may have exerted considerable control on the $\delta^{18}O$ records (Alt *et al.* 1985).

The Neoglacial is also coincident with the establishment of many of the ice shelves along northern Ellesmere Island coast, although in some cases, growth may have been initiated as early as 4 ka BP (Lyons & Mielke 1973; Stewart & England 1983). It is also roughly accordant with the decline in regional peat accumulation, although the interpretation and significance of this proxy record is uncertain (cf. Ovenden 1988; Bradley 1990; LaFarge-England *et al.* 1991; Gajewski *et al.* 1995).

Results from the study of varved sediments in Lake C2 (Fig. 1, site 3), demonstrate an abrupt shift from high sedimentation rates between 3.3-3.1 ka BP, to below average sedimentation from 3.0-2.3 ka BP, suggesting a climatic cooling (Lamoureux & Bradley 1996). There are three important considerations in interpreting the significance of these data: 1) that the basin is thought to have emerged from the sea ~2.7 ka BP, and thus sedimentation prior to this potentially reflects different hydrological and sedimentological controls than in the post-emergent period, 2) that sedimentation in Lake C2 is most strongly correlated with temperatures 600 m above Alert (Hardy *et al.* 1996), reflecting the topographic distribution of snow and glaciers feeding the lake, and 3) that the lake is considered to have remained perennially ice-covered throughout the late-Holocene (Hattersley-Smith *et al.* 1970; Lamoureux & Bradley 1996). Between 2.3-1.2 ka BP sedimentation in Lake C2 was slightly above average (although there was a high inter-annual variability), which is interpreted as indicating warmer conditions (Lamoureux & Bradley 1996). In contradiction of this, the post-1.9 ka BP decline in diatom abundance in the Appleby Lake core (Fig. 2.6) suggests an even further climatic deterioration, this time affecting the lower Lake Hazen Basin itself (snowline below 450 m). Diatom abundances are approximately equal to those recorded in the period (~5-4 ka BP) prior to the mid-Holocene rise. Cooler conditions are also recorded along western Greenland (Kelly 1980;Weidick *et al.* 1990) and generally conform to an overall declining $\delta^{18}O$ record in the Agassiz, Camp Century and Devon ice cores (Dansgaard *et al.* 1982; Koerner 1989). It may be useful, however, to note that detrended $\delta^{18}O$ curves for the Camp Century and Devon ice cores (*in* Lamoureux & Bradley 1996) are greater than normal (more +ve) from ~2000-1000 BP, suggesting an increase in southerly derived moisture. An increased southerly air flow is also supported by the exotic pollen records from the Agassiz ice cap which peak from 1.1-1.5 ka BP, and are generally highest from 3.1 ka BP to present (Bourgeois 1986). The differing Lake C2 and Hazen records may be reconciled by interpreting this period as one of generally increased snowfall (higher albedo), accompanied by a delay in the onset of summer melt, which results in a more continuous lake ice cover and reduced transmission of light into the lake.

2.5 CONCLUSIONS

This study presents a Holocene paleoenvironmental record from the continental interior of northeastern Ellesmere Island based on the diatom assemblages in cores from four lakes. The extent and persistence of a summer lake ice cover appears to be the greatest limitation to diatom growth in this region. Consequently, paleolimnological records of diatom abundance serve as a proxy ice-cover record, whereby increased productivity is linked with more open-water conditions (cf. Smol 1988). Changes in diatom abundance within the four lakes studied, range from 0 to 10^9 valves/gram dry sediment, indicating that considerable environmental changes have occurred during the Holocene. The following paleoenvironmental reconstruction is proposed.

During deglaciation, much of the eastern and central Lake Hazen Basin and Hazen Plateau remained ice covered from ~8-6 ka BP. Retreat was subsequently rapid, and by 5 ka BP, glaciers had retreated to the north shore of Lake Hazen. To the south, plateau ice caps persisted on the higher Hazen Plateau. Diatom remains are generally sparse, or absent, within much of the lower lake sediments dating from ~7.6-5 ka BP, suggesting extensive perennial ice cover. Climate records from elsewhere on the Hazen Plateau (LaFarge-England *et al.* 1991) and throughout the Arctic (Bradley 1990), point to warmer conditions during this time, suggesting that strong local climatic controls relating to the proximity of glaciers may have been affecting the study lakes. The first significant rise in diatom concentrations occurs between ~5 and 4 ka BP, coeval with the retreat of glaciers from Lake Hazen. While a proportionally similar rise in diatom abundance is noted in each of the four basins, the chronostratigraphic link between the four basins is provisional (Fig. 2.6). Between 4 and 3 ka BP, diatom abundance peaks in the Brainard Lake core, and is equally high in Appleby Lake. The absence of chronological control, and uncertainties in the extent of preservation of the surface sediments from Hart and Whisler lakes preclude recognition of this event. This 4-3 ka BP rise is considered to mark the warmest conditions throughout Lake Hazen Basin and Hazen Plateau recorded in this study. This corroborates regional archaeological records that show a high incidence of Independence I cultural sites dating between 4 and 3 ka BP (Sutherland 1991b; Dick *et al.* 1994). Proxy records from elsewhere in the High

Arctic also record warmer conditions during this period, although many from coastal areas suggest highest warming occurred prior to 4 ka BP, declining sharply around 3.5 ka BP, or earlier (Smol 1983; Blake *et al.* 1992). Discrepancies between the records highlight the regional topoclimatic variability (Edlund & Alt 1989; Alt & Maxwell 1990), the effect of the North Open Water polynya (Koerner 1979; Müller *et al.* 1980), and possible synoptic variations (Alt 1983, 1985). Post 3 ka BP, diatom abundances decline sharply in Brainard Lake, to lower, though still moderate levels of productivity. However, diatom abundances remain high (10^9 valves/gram dry sediment) in Appleby Lake until 1.9 ka BP, after which they decline. Differences between the two lake records highlight the strong topoclimatic variations seen in the regions today, whereby lakes higher on Hazen Plateau experience markedly cooler conditions and more persistent ice cover than lakes situated in the lower Lake Hazen Basin. The decline in diatom abundance in Brainard Lake is thus considered to mark an overall decline in the regional snowline. This correlates with both the growth of ice shelves off northern Ellesmere Island and the extensive Neoglacial advance recorded throughout western Greenland and southeastern Ellesmere Island. Around 1.9 ka BP, diatom abundances decline markedly in the Appleby Lake core, bottoming out at 1.5 ka BP, and then maintaining generally lower levels through the remaining, upper 10 cm of sediment. This decline in productivity suggests an even further cooling, this time affecting lakes in Lake Hazen Basin, in addition to those on the surrounding Hazen Plateau. Detailed studies of varved lake sediments from northern Ellesmere Island have suggested that increased varve thickness between 2.3-1.2 ka BP relates to increased melting (Lamoureux & Bradley 1996). Whether this reflects increased precipitation or increased temperature is uncertain. However, such regional differences are both valid and instructive, and thus serve to demonstrate the inherent variability of paleoenvironmental records from the High Arctic.

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2.7 APPENDIX

Appendix 2.1 Diatom Species list and Taxonomic Authority

[A = Appleby Lake; B = Brainard Lake; H = Hart Lake; W = Whisler Lake]

Achnanthes flexella (Kütz)Brun [B, H]
Achnanthes joursacense Héribaud [W]
Achnanthes laevis Østrup [B]
Achnanthes minutissima Kützing [A,B,H,W]
Achnanthes modestiformis W
Achnanthes oestrupii (Cleve-Euler)Hustedt [W]
Actinoptychus sp. ?*undulatus* (Bailey)Ralfs. [H] {Tertiary form}
Amphora delicatissima Krasske [B]
Amphora fogediana Krammer [B,H,W]
Amphora inariensis Krammer [B,H,W]
Amphora lybica Ehrenberg [A,B,H,W]
Amphora pediculus (Kützing)Grunow ex A.Schmidt [B,H,W]
Asterionella sp. [W]
Aulocoseira ambigua (Grunow in Van Heurck)Krammer [W]
Aulocoseira distans (Ehrenberg)Simonsen [H,W]
Aulocoseira granulata (Ehrenberg)Simonsen [W]
Aulocoseira italica (Ehrenberg)Simonsen [W]
Aulocoseira subarctica (O. Müller)Haworth [A,W]
Caloneis alpestris (Grunow)Cleve [W]
Caloneis bacillum (Grunow)Cleve [W]
Caloneis silicula (Ehrenberg)Cleve [H,B]
Cocconeis placentula var. *lineata* (Ehrenberg)Van Heurck [H,W]
Cocconeis placentula Ehrenberg [H,W]
Cocconeis pseudothumensis Reichardt [W]
Cyclotella bodanica var. *affinis* Grunow [W]
Cyclotella bodanica var. *bodanica* Grunow [A,H,W]
Cyclotella bodanica var. *lemanica* (O.Müller ex Schröter) Bachmann [A,H,W]
Cyclotella glomerata Bachman [W]
Cyclotella meneghiniana Kützing [W]
Cyclotella radiosa (Grunow)Lemmermann [W]
Cymbella amphicephala Nägeli [B]
Cymbella cesatii (Rabenhorst)Grunow in A.Schmidt *et al* [W]
Cymbella cistula (Ehrenberg in Hemprich & Ehrenberg) Kirchner [W]
Cymbella cuspidata Kützing [W]
Cymbella designata Krammer [W]
Cymbella elginensis Krammer {=Encyonema elginensis (Krammer)Mann in Round, Crawford & Mann [W]
Cymbella incerta (Grunow)Cleve [B,W]
Cymbella microcephala Grunow in Van Heurck [A]
Cymbella minuta Hilse in Rabenhorst {=Encyonema minutum (Hilse in Rabenhorst)Mann in Round, Crawford & Mann [A,B,H,W]
Cymbella naviculiformis var. *linearis* Foged [B]
Cymbella sp. [cf. *obtusiuscula* (Kützing)Grunow] [B]
Cymbella perpusilla Cleve [B]
Cymbella reichardtii Krammer {=Encyonema reichardtii (Krammer) Mann in Round, Crawford & Mann [B]
Cymbella silesiaca Bleisch {=Encyonema silesiacum (Bleisch in Rabenhorst)Mann in Round, Crawford & Mann [A,B,H,W]
Cymbella subaequalis Grunow in Van Heurck [W]
Denticula kuetzingii Grunow [B,W]
Diatoma hiemale var. *mesodon* (Ehrenberg)Kirchner [B,W]
Diatoma moniliformis Kützing [A,B]
Diatoma tenue Agardh [W]
Diploneis oculata (Brébisson)Cleve [W]
Diploneis ovalis (Hilse)Cleve [H]
Diploneis pseudovalis Hustedt [H]
Distephanus speculum? (in Tynni 1982, plate XVII, #20) [H] {Tertiary form}
Eunotia diodon Ehrenberg [W]
Eunotia praerupta Ehrenberg [H]
Fragilaria brevistriata Grunow {=*Pseudostaurosira brevistriata*} [A,B,H,W]
Fragilaria capucina var. *vaucheriae* (Kützing)Lange-Bertalot [B]
Fragilaria construens (Ehrenberg)Grunow {=*Staurosira construens* (Ehrenberg)Williams & Round} [A,B,H,W]
Fragilaria construens var. *pumila* Grunow in Van Heurck [W]
Fragilaria construens var. *venter* (Ehrenberg)Grunow in Van Heurck {=*Staurosira construens* var. *venter* (Ehrenberg)Hamilton in Hamilton Poulin Charles and Angell} [A,B,H,W]
Fragilaria crotonensis Kitton [W]
Fragilaria lapponica Grunow in Van Heurck {=Staurosirella lapponica (Grunow in VanHeurck)Williams & Round} [B,H]
Fragilaria leptostauron (Ehrenberg)Hustedt {=Staurosirella lapponica (Grunow in VanHeurck)Williams & Round} [W]

Fragilaria pinnata Ehrenberg {=Staurosirella pinnata (Ehrenberg)Williams & Round} [A,B,H,W]

Fragilaria pinnata var. *accuminata* A.Mayer [A,B,W]

Fragilaria pinnata var. *intercedens* (Grunow in Van Heurck)Hustedt {=*Staurosirella pinnata* var. *intercedens* (Grunow in Van Heurck)Hamilton in Hamilton Poulin Charles and Angell. [A,B,H,W]

Fragilaria pinnata var. *lancettula* (Schumann)Hustedt in Schmidt *et al* [A,H,W]

Fragilaria pseudoconstruens Marciniak {=*Pseudostaurosira pseudoconstruens* (Marciniak)Williams & Round} [A,B,H,W]

Gomphonema sp. [cf. *angustatum* Agardh] [W]

Gomphonema angustatum (Kützing)Rabenhorst [B,H,W]

Gomphonema bohemicum Reichelt & Fricke [W]

Gomphonema gracile Ehrenberg [W]

Gomphonema olivaceoides Hustedt [B]

Gomphonema sp. [cf. *parvulum* (Kützing)Kützing] [H]

Gyrosigma kützingii (Grunow)Cleve [W]

Hantzschia amphioxys (Ehrenberg)Grunow [W]

Navicula aquaedurae Lange-Bertalot B

Navicula bacillum Ehrenberg {=Sellaphora bacillum (Ehrenberg) Mann in Round, Crawford & Mann [W]

Navicula sp. [cf. *bahusiensis* (Grunow in Van Heurck) Grunow] [B]

Navicula capitata var. *hungarica* (Grunow)Ross [W]

Navicula sp. [cf. *cari* Ehrenberg] [B]

Navicula sp. [cf. *cincta* var. *heufleri* (Grunow)Grunow in Van Heurck] [B,H]

Navicula cryptocephala Kützing [A,B,H,W]

Navicula cuspidata Kützing [B]

Navicula digitoradiata (Gregory)Ralfs in Pritchard [B,H]

Navicula disjuncta Hustedt {=Sellaphora disjuncta (Hustedt) Mann in Round, Crawford & Mann [W]

Navicula sp. [cf. *eidrigiana* Carter] [W]

Navicula elginensis (Gregory)Ralfs in Pritchard [B,H]

Navicula halophila (Grunow)Cleve [A,B]

Navicula hilliardii Manguin [B,H,W]

Navicula laevissima Kützing {=Sellaphora laevissima (Kützing) Mann [A,B,H,W]

Navicula laevissima var. *perhibita* (Hustedt)Lange-Bertalot [B,W]

Navicula lanceolata (C.Agardh)Ehrenberg [W]

Navicula lundii Reichardt [B]

Navicula minuscula Grunow in Van Heurck [B]

Navicula modica Hustedt [W]

Navicula sp. [cf. *molestiformis* Hustedt] [B,H]

Navicula mutica Kützing {=*Luticola mutica* (Kützing)Mann in Round, Crawford & Mann} [W]

Navicula sp. [cf. *oppugnata* Hustedt] [W]

Navicula phyllepta Kützing [B]

Navicula placentula (Ehrenberg)Kützing [W]

Navicula porifera var. *opportuna* Hustedt [W]

Navicula pseudanglica Cleve-Euler [B]

Navicula pseudanglica var. *signata* (Hustedt)Lange-Bertalot [B]

Navicula pseudoscutiformis Hustedt {=Cavinula pseudoscutiformis (Hustedt) Mann & Stickle in Round, Crawford & Mann [B,H,W]

Navicula pupula Kützing {=Sellaphora pupula (Kützing) Mereschkowsky in Round, Crawford & Mann [A,B,H,W]

Navicula radiosa Kützing [A]

Navicula reinhardtii Grunow in Van Heurck [B]

Navicula rhyncocephala Kützing [W]

Navicula schoenfeldii Hustedt [B,W]

Navicula scutelloides W.Smith ex Gregory [W]

Navicula similis Krasske [H]

Navicula soehrensis Krasske [W]

Navicula subminuscula Manguin [B]

Navicula trivialis Lange-Bertalot [W]

Navicula tuscula Ehrenberg {=Aneumastus tusculus (Ehrenberg) Mann & Stickle in Round, Crawford & Mann [W]

Navicula vulpina Kützing [A,B]

Neidium ampliatum (Ehrenberg)Krammer [B]

Neidium iridis (Ehrenberg)Cleve [B]

Nitzschia denticula Grunow [B]

Nitzschia dissipata var. *media* (Hantzsch)Grunow in Van Heurck [B]

Nitzschia inconspicua Grunow [B,W]

Nitzschia paleacea (Grunow in Cleve & Grunow)Grunow in Van Heurck [B]

Nitzschia perminuta (Grunow)M.Peragallo [B,W]

Nitzschia pusilla (Grunow)Lange-Bertalot [B]

Nitzschia recta Hantzsch ex Rabenhorst [B]

Nitzschia rosenstockii Lange-Bertalot [W]

Opephora sp. [cf. *olsenii* Møller] [A,H,W]

Pinnularia balfouriana Grunow ex Cleve [H,W]

Pinnularia borealis var. *rectangularis* Carlson [B]

Pinnularia microstauron var. *brebissonii* (Kützing)Hustedt [B,W]

Pinnularia viridis (Nitzsch)Ehrenberg [A]

Pinnularia sp. [W]

Stauroneis anceps Ehrenberg [B]

Stauroneis phoenicenteron (Nitzsch)Ehrenberg [B]

Stephanodiscus alpinus Hustedt [H,W]

Stephanodiscus binatus (Håkansson & Kling) [A,W]

Stephanodiscus hantzschii Grunow in Cleve & Grunow [W]

Stephanodiscus minutulus (Kützing)Round [W]

Stephanodiscus niagarae Ehrenberg [W]

Stephanodiscus parvus Stoermer & Håkansson [A,W]

Surirella sp. [W]

Synedra ulna (Nitzsch)Ehrenberg [W]

Tabellaria flocculosa (Roth)Kützing [H,W]

INTERPRETATION OF DIAMICTIC SEDIMENTS WITHIN HIGH ARCTIC LACUSTRINE CORES: EVIDENCE FOR LAKE ICE-RAFTED DEPOSITION

3.1 INTRODUCTION

Sixteen small lake basins on northern Ellesmere Island in the Canadian High Arctic were cored as part of a study of the glacial history and paleoenvironmental record (Fig. 3.1). The uppermost sediment (50-150 cm) in each of the lakes is a massive, clayey-silt. Underlying this is one of two contrasting facies. One is a massive, gravel-rich to gravel-poor diamict. This occurred in six lakes, and the maximum recovery was 2.1 m. The other lakes displayed variable assemblages of sorted sediment (up to 1.8 m) including sand and gravel (much of which was current-bedded), proximal-distal fining rhythmites, and thin (<0.4 m) sections of gravel-rich diamict with sand and gravel interbeds. The objectives of this study were to resolve the genesis of the diamicts from the two facies. Determining whether they represented tills, proglacial lake sediments or some other type of deposit has important implications for the regional glacial history.

Lacustrine sediments in non-glaciated Arctic catchments are generally fine grained. Deposition of discrete sand lenses and occasional gravel and pebbles are linked to a variety of processes, including slumping and turbidity currents (Gilbert & Church 1983; Lemmen *et al.* 1988; Fredskild 1995). Other mechanisms of transporting coarse sediments within these environments relate to the persistence of a thick and often "quasi-permanent" ice floe. Examples include, alluvial deposition, slush flows and slumping onto the ice floe prior to initial breakup (Barnes 1960; Allan *et al.* 1987; Doran 1993; Retelle & Child 1996), adfreezing to the ice floe within the littoral zone (Shilts & Dean 1975), the formation and breakup of frazil and anchor ice (Tsang 1982; Osterkamp & Gosink 1983), and the physical plowing of sediments by the ice floe (Dionne 1979). Periglacial processes also contribute to basinward movement of coarser sediments (Livingstone *et al.* 1958; Mackay 1967; Shilts & Dean 1975; Doran 1993). Within smaller lake basins without any sizeable fluvial input, gravel-rich facies have not generally been recognized, other than those attributed to glacial deposition (till and/or glaciolacustrine; Blake 1981, 1987; Blake *et al.* 1992), or marine isolation basins (Blake 1982; Retelle 1986; Bennike 1995).

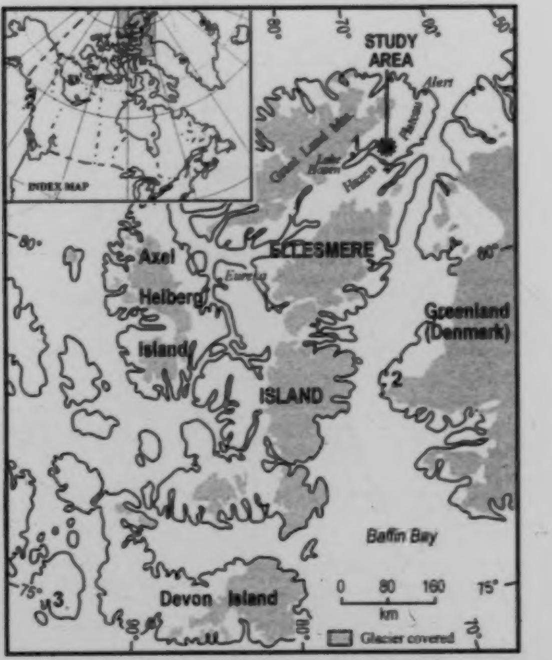
INDEX MAP
STUDY AREA
Alert
Grant Land Mts.
Lake Hazen
Hazen
Plateau
ELLESMERE
ISLAND
Axel
Heiberg
Island
Eureka
Greenland
(Denmark)
2
3
Baffin Bay
Devon Island
80°
75°
0 80 160
km
Glacier covered

Figure 3.1 Study area located on northeastern Ellesmere Island, within the Canadian High Arctic. Bold-faced numbers are sites referred to in the text.

3.1.1 Study Area

The Lake Hazen region is a key biological oasis within the Canadian Arctic, yet there has been surprisingly little research conducted on its paleoenvironmental record. Previous research had proposed that large areas of the encircling Hazen Plateau remained ice-free during the last glacial maximum (LGM; England 1976, 1978). Extant lake basins within and beyond the proposed ice limit of the LGM were cored to test this hypothesis and provide additional, important evidence on Holocene paleoenvironmental change.

The area is characterised by three physiographic regions, including the Grant Land Mountains which trend SW-NE along the Lake Hazen Fault Zone, north of Lake Hazen (Fig. 3.1). Mountains reach 2665 m above sea level (asl) and support a large ice cap, from which trunk glaciers descend to between 500 and 250 m asl along Lake Hazen. Parallelling the mountains is Lake Hazen Basin. Lake Hazen is 70 km long, 290 m deep, 540 km^2, and 158 m asl. East of Lake Hazen Basin, the northward dipping Hazen Plateau rises from 300 to 1300 m asl, contacting the sea along spectacular cliff-bounded fiords. Surficial materials are predominantly bedrock (calcareous and dolomitic sandstone and mudrock; Trettin 1994) and a patchy till veneer, although local accumulations of glaciolacustrine sediment up to 40 m thick occur.

Climatically, the region is described as a polar desert (Bovis & Barry 1974), with a mean annual temperature of -20°C, and a mean annual precipitation of less than 150 mm water equivalent (Thompson 1994). Owing to its intermontane setting and reflection of incident radiation along the SW-NE trending aspect of the Grant Land Mountains and off perennially ice-covered Lake Hazen, Lake Hazen Basin becomes a "thermal oasis" in the brief summer. Temperatures can exceed 18°C. Cold-air drainage and severe temperature inversions account for mean winter monthly temperatures (-40°C) that are 6 to 10°C colder than at Alert, on the northeast coast (Fig. 3.1; Thompson 1994).

Numerous small, shallow lakes and ponds are found throughout the study area (Fig. 3.2), almost all of which have bedrock-controlled outlets. Maximum lake ice thicknesses vary from 1.9 to 3.2 m, similar to other High Arctic lakes (Doran 1993; Heron & Woo 1994). There is a strong altitudinal temperature gradient in the region, leading to increased lake ice thicknesses on the plateau ($\geq$2.2 m), compared to $\leq$2 m on lakes in Lake Hazen Basin. This gradient also accounts for the delayed snowmelt and breakup of the lake ice on the plateau,

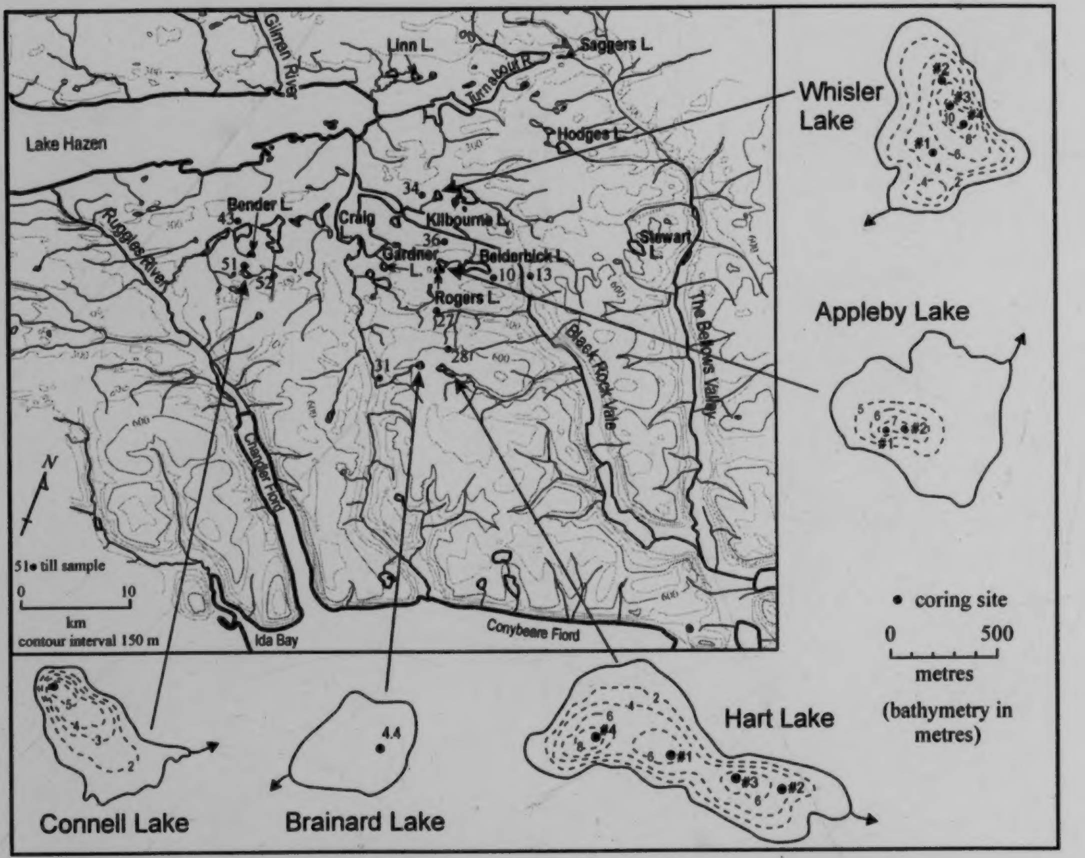
Gilman River
Linn L.
Saggers L.
Turnabout R.
Whisler Lake
Lake Hazen
Hodges L.
34
Bender L.
43
Craig
Kilbourne L.
36
Stewart L.
Ruggles River
51
52
Gardner L.
Belderbick L.
10
13
Rogers L.
27
Appleby Lake
28
31
Black Rock Vale
The Bellows Valley
Chandler Fiord
N
51• till sample
0
10
km
contour interval 150 m
Ida Bay
Conybeare Fiord
• coring site
0
500
metres
(bathymetry in metres)
Hart Lake
Connell Lake
Brainard Lake
#1
#2
#3
#4
4.4

Figure 3.2 Topographic map highlighting the regional setting of the lakes cored as part of this study. Numbers indicate surficial till collection sites. Detail of five of the study lakes show bathymetry in metres and location of coring sites.

e.g., at the beginning of August 1994, *Hart* and *Brainard* lakes (unofficial lake names are indicated in italics at their first mention, Fig. 3.2) retained greater than 90 and 70% ice cover, respectively, while all of the lakes in Lake Hazen Basin (excluding Lake Hazen) had cleared by at least mid-July.

Table 3.1 summarizes characteristics of the main study lakes.

3.1.2 Core Sedimentology

Sediment core logs representative of the two facies underlying the clayey-silt are shown in Figure 3.3. A massive, gravel-rich to gravel-poor diamict was found in a total of 15 cores from Appleby, Biederbick, Brainard, Hart, *Stewart*, and *Whisler* lakes (Figs. 3.2 and 3.3). The diamicts are characterised by a sandy-silt matrix containing varying proportions of gravel (~2 - 30% by weight) up to 6 cm in diameter (Fig. 3.4). As the inner core barrel diameter is only 6.6 cm, it is likely that contact with these and larger clasts halted core penetration. Multiple cores from the same lake basin reveal a compositional heterogeneity (% gravel), although there is an overall similarity in the sedimentology of diamicts from the six lake basins (Fig. 3.4). Many of the clasts exposed along the cut face have a high dip (Fig. 3.4). The upper contact between the diamict and overlying clayey-silt is irregular, but sharp.

Sediments underlying the clayey-silt belonging to the second facies assemblage consisted of 1) alternating sequences of diamict, sand and gravel, some of which are current-bedded (*Bender*, *Connell* and *Gardiner* lakes), 2) massive sand and gravel (*Hodges*, *Linn* and *Saggers* lakes), or 3) rhythmically-bedded sand and/or silt and clay which displays proximal-distal fining (Kilbourne Lake; Fig. 3.3). Current-bedded stratification is principally *Type A* climbing ripple cross-stratification, although *Type B* and planar cross-stratified beds are also present. In lakes displaying alternating beds of diamict and sorted sediment, thin-sections of the upper diamict contacts reveal diapiric structures and occasional, small rip-ups suggestive of erosion and transport by traction. The upper diamict contacts also show an accumulation (<2 mm) of fines, indicating a period of relatively quiescent sedimentation. Sections of the diamict within Bender, Connell and Gardiner lakes, display weak sorting, and discontinuous thin interbeds of gravel and sand (Fig. 3.3).

Coal is found throughout Lake Hazen Basin within the till veneer, and often blankets the shores of lakes. Eocene coal beds run parallel to the north shore of Lake Hazen (Miall

Table 3.1 Morphometric parameters and other characteristics of the main study lakes.

Lake	Elevation (m asl)	max. Depth (m)	Surface Area (km^2)	Volume ($x 10^6 m^3$)	Catchment Area (km^2)	D_{25}[1]	D_{50} (m)	D_{75}	% of Basin Shallower than 2 m	% of Basin Shallower than 3 m
Appleby	447	7.70	0.488	n.d.[2]	3.269					
Bender	300	7.25	0.249	1.047	1.192	4.44	2.88	1.02	39	52
Brainard	632	4.40	0.176	n.d.[2]	2.115					
Connell	310	5.75	0.303	0.889	2.102	2.86	1.97	1.20	51	77
Hart	621	8.50	0.677	2.936	4.802	4.18	2.22	1.20	45	60
Whisler	245	11.40	0.311	1.447	1.466	4.72	2.43	1.10	42	58

[1] Quartile depths: indicates depth at which cumulative percentage area of lake lies below (Håkanson, 1981)

[2] There is insufficient bathymetric control available to determine volume and quartile depth measurements from Brainard and Appleby Lake.

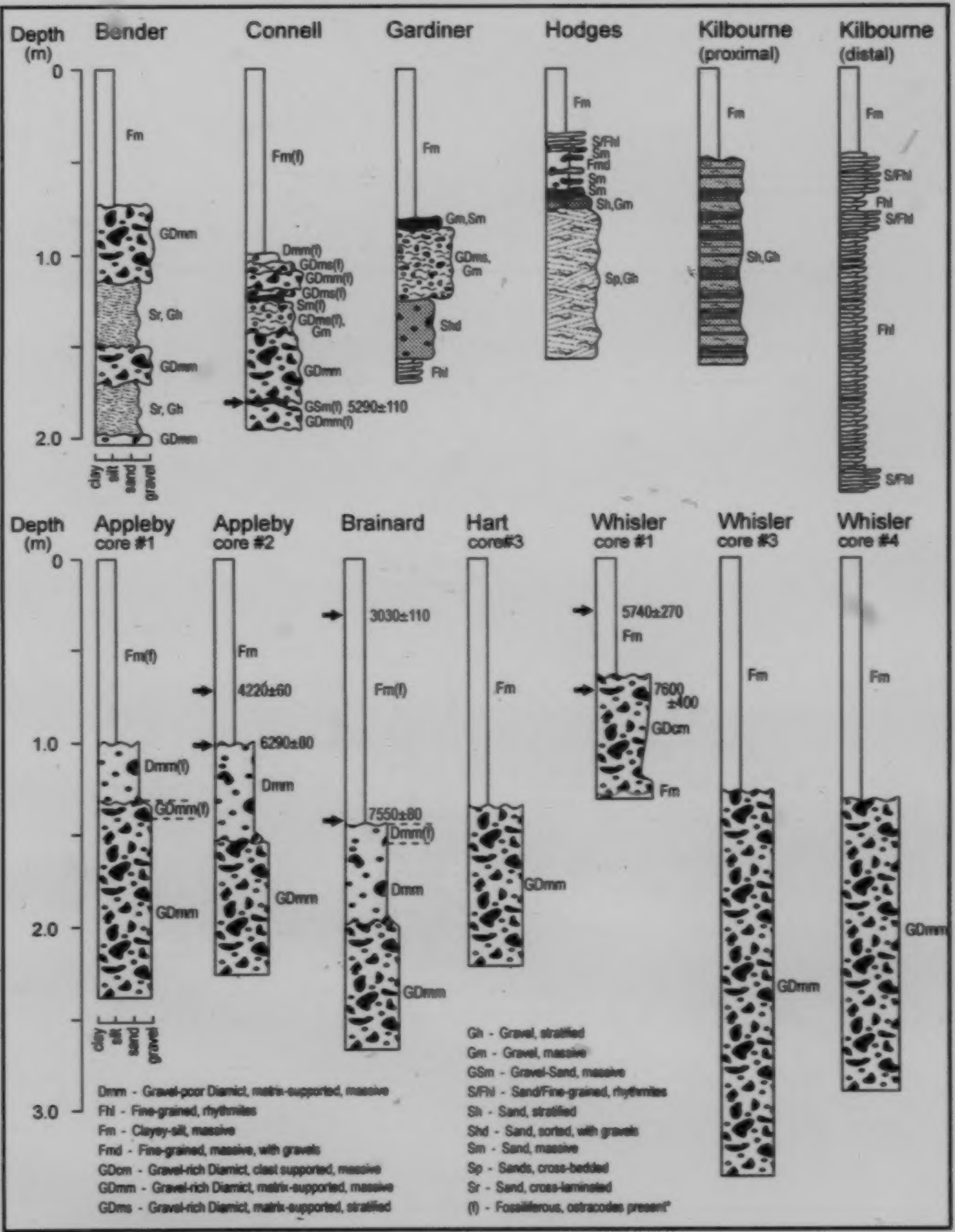
Depth (m)
0
1.0
2.0
Bender
Fm
GDmm
Sr, Gh
GDmm
Sr, Gh
GDmm
clay
silt
sand
gravel
Connell
Fm(f)
Dmm(f)
GDms(f)
GDmm(f)
GDms(f)
Sm(f)
GDms(f), Gm
GDmm
GSm(f) 5290±110
GDmm(f)
Gardiner
Fm
Gm,Sm
GDms, Gm
Shd
Fhl
Hodges
Fm
S/Fhl
Sm
Fmd
Sm
Sm
Sh,Gm
Sp,Gh
Kilbourne (proximal)
Fm
Sh,Gh
Kilbourne (distal)
Fm
S/Fhl
Fhl
S/Fhl
Fhl
S/Fhl
Depth (m)
0
1.0
2.0
3.0
Appleby core #1
Fm(f)
Dmm(f)
GDmm(f)
GDmm
clay
silt
sand
gravel
Appleby core #2
Fm
4220±60
6290±80
Dmm
GDmm
Brainard
3030±110
Fm(f)
7550±80
Dmm(f)
Dmm
GDmm
Hart core#3
Fm
GDmm
Whisler core #1
5740±270
Fm
7600 ±400
GDcm
Fm
Whisler core #3
Fm
GDmm
Whisler core #4
Fm
GDmm
Dmm - Gravel-poor Diamict, matrix-supported, massive
Fhl - Fine-grained, rhythmites
Fm - Clayey-silt, massive
Fmd - Fine-grained, massive, with gravels
GDcm - Gravel-rich Diamict, clast supported, massive
GDmm - Gravel-rich Diamict, matrix-supported, massive
GDms - Gravel-rich Diamict, matrix-supported, stratified
Gh - Gravel, stratified
Gm - Gravel, massive
GSm - Gravel-Sand, massive
S/Fhl - Sand/Fine-grained, rhythmites
Sh - Sand, stratified
Shd - Sand, sorted, with gravels
Sm - Sand, massive
Sp - Sands, cross-bedded
Sr - Sand, cross-laminated
(f) - Fossiliferous, ostracodes present*

Figure 3.3 Sediment core logs. Upper log diagrams highlight the variety of sediments found in the sorted sediment and mixed diamict facies assemblage. Lower core logs are from those lakes with the primary diamict facies assemblage. Note, the presence of ostracodes was only assessed in Appleby core #1, Brainard, and Connell lakes, and the diamict of Whisler core #3. Chronology of different cores is provided by radiocarbon dates (Table 3.2).

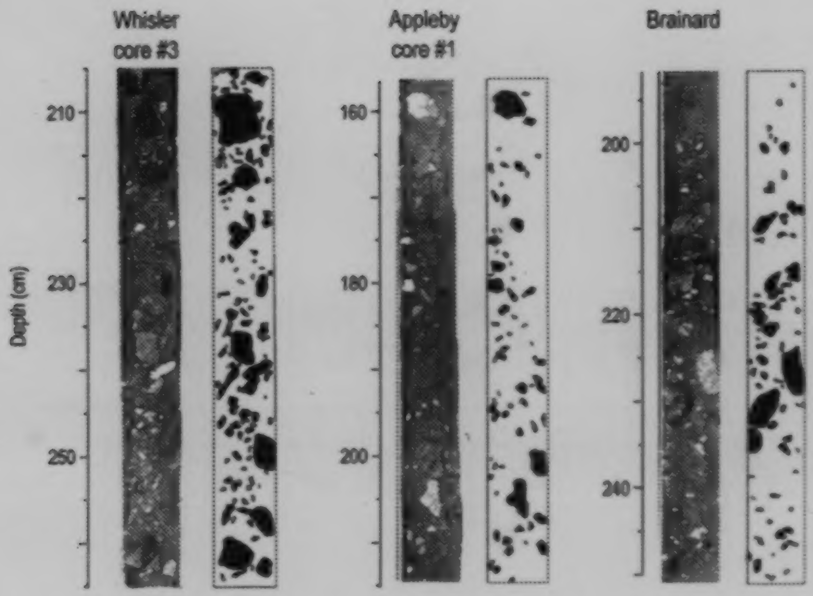
Whisler core #3
Appleby core #1
Brainard
Depth (cm)
210
230
250
160
180
200
200
220
240

Figure 3.4 Photographs of sections of the gravel-rich diamict (GDmm), from three of the study lakes. Paired tracings of clasts highlights their occurrence and preferred near-vertical orientation. Disturbance in the Brainard core at 214 cm is a core segment break.

1979). Large deposits of coal occur within the catchments of Bender, Biederbick, Hart and Brainard lakes. Within current-bedded sediments, coal occurs abundantly in the lee side laminae of ripples, and as discrete lenses (up to 23 mm thick) within other sorted gravel and sand. Fine coal is also found within the matrix of the diamicts and as angular gravels, yet never occurs as a discrete lens or accumulation.

3.2 METHODS

3.2.1 Sediment Cores

Sediment cores were extracted in June 1993, using a percussion corer (Reasoner 1993). In order to penetrate further into the diamict, a vibracorer (Smith 1992) was used in May 1994. Coring sites were located within the deepest basin of each lake, as determined by spot depths from holes drilled through the ice floe. Logistical constraints generally limited the recovery to only a single core from each lake. In an attempt to assess inter-basinal variations, four cores (using both coring systems) were removed from Whisler and Hart lakes, and two cores from Appleby and Stewart lakes. Cores were capped and sealed in the field, shipped south, and then frozen and split lengthwise with a diamond saw at Core Laboratories, Calgary. The cut face was washed, while frozen, to remove any contamination from the splitting process. The use of a core catcher in both coring systems, and the freezing of the cores, resulted in variable amounts of downwarping, or coning, of the upper clayey-silt (2 - 8 cm). There is an obvious progression to minimal downwarping (<1 cm) towards and within the diamict. Up to 2 cm of downwarping is observed in some of the sorted sediments.

3.2.2 Clast Fabric Analysis

Clast fabrics have been used by researchers to characterize different types of till and ice-rafted deposits (Lawson 1979, Domack & Lawson 1985, Dowdeswell & Sharp 1986). Clast fabrics of ice-rafted diamicts have a random clast azimuth and a higher percentage of clast dips >45° than do tills and other glaciogenic diamicts (Domack & Lawson 1985). Clast fabrics were thus determined from the diamicts of Appleby Lake core #1 and #2, Brainard Lake, Connell Lake, and Whisler Lake core #1 and #3.

The cohesion of the diamicts (owing to the abundance of fines), allowed the progressive dissection of core segments, using picks and tweezers to pry the material apart. All clasts with an *a:b* axial ratio greater than 1.5, and an *a*-axis length greater than 8 mm, were measured. Orientation and dip were measured with a protractor and Brunton compass (azimuth was measured relative to the plane of the vertical cut face, clockwise from the 0° left margin). Paired sections of larger clasts split by the lengthwise cutting of the cores were matched, and adjusted for ~4 mm of removal by the saw. To avoid effects of disruption by the coring process, clasts within 1 cm of the coring tube walls were excluded. The data were plotted on Schmidt equal-area lower hemisphere projections contoured at 1 standard deviation intervals using the method of Kamb (1959). Clasts from the diamicts were inspected for evidence of striations and carbonate precipitate coatings, and comparisons were made with those from 10 surface till samples (#10, 13, 27, 28, 31, 34, 36, 43, 51, 52; Fig. 3.2).

3.2.3 Particle-size

Particle-size determinations were made on the diamicts, overlying sediment, and the 10 surface tills. Gravel (>2 mm) and sand (2-0.063 mm) contents were determined by dry sieving, while the silt/clay content (0.063-0.002 and <0.002 mm, respectively) was analyzed using a Micrometrics® 5100 SediGraph.

Particle-size distributions of the diamicts and tills were polymodal, so the technique of Gaussian component analysis (Sheridan *et al.* 1987; Sharp *et al.* 1994) was used to separate each sample distribution into its constituent sub-populations. The mean size, standard deviation and fraction of the total sample of each sub-population were calculated, as was the magnitude of the unexplained residuum. Three to six Gaussians were required to accurately describe the diamicts, leaving unaccounted residua of 1.81 - 8.38 %.

3.2.4 Subfossil Remains

Microscopic scans of diamicts and sorted sediment from several cores revealed a rich subfossil assemblage (ostracodes, diatoms, chrysophyte stomatocysts, chironomid head capsules, cladocera egg cases, seeds and other aquatic and terrestrial plant material). The presence of these cannot of itself be considered proof of a non-glacial origin, as numerous studies have clearly demonstrated the capacity for glaciers to transport paired marine pelycopod shells, foraminifera, diatoms and ostracodes (Boulton 1970; Mangerud *et al.* 1981; Petersen 1983; Nielsen *et al.* 1986; Hicock & Dreimanis 1992). Rather than simple presence, such factors as habitat affinity and assemblage composition may be important to distinguish glacially "mixed" populations from unaltered, *in situ* populations. The assemblages of different subfossils were thus documented in cores representative of the two facies, and from the 10 surface tills.

3.2.5 Ostracodes

Ostracodes (class Crustacea) are a calcitic microfauna that live at the sediment-water interface in a wide variety of freshwater and marine environments. Their distribution is related to a variety of ecological parameters, including pH, temperature and habitat (Delorme 1991), but they have rarely been reported from high Arctic regions (Røen 1981). Ostracodes were hand-picked under a microscope from >63 μm sieved subsamples of the diamict from Whisler core #3, the diamict and overlying clayey-silt of Appleby core #2 and the Brainard Lake core, and the diamict, sorted sediment and overlying clayey-silt of the Connell Lake core. The relative proportions of articulated (both shells attached) and single shells was determined on several samples prior to any pretreatment. Proportions of articulated shells are likely underestimates, as even minimal handling resulted in breakage of shells.

3.2.6 Diatoms and Chrysophyte Stomatocysts

Diatoms (class Bacillariophyceae) and chrysophytes (classes Chrysophyceae and Synurophyceae) are diverse groups of algae, found within a wide range of aquatic environments. Diatoms possess a cell wall (frustule) of biogenic silica, the morphology of which is unique and identifiable to the species level. Chrysophytes produce unique species-specific siliceous resting stages known as stomatocysts or statospores, but they are more difficult to identify due to their smaller size, and taxonomic uncertainties. Remains of both diatoms and chrysophyte stomatocysts are often well preserved in Arctic lake sediments, and species distribution can be linked to environmental parameters, including habitat (Smol 1983, 1987, 1988; Duff & Smol 1989).

Studies by the author (unpublished data) have shown that within lake basins on Hazen Plateau, the littoral and benthic diatom populations are strongly dissimilar. Furthermore, scrapings of clasts from the lakes show a high diversity and abundance of epilithic species (attached to rocks). Thus, in an attempt to further resolve the genesis of the diamicts and possible origin of the clasts, scrapings were taken from clasts within the diamicts of Brainard, Appleby and Whisler lakes. These lakes were chosen because detailed diatom studies had previously been made of the overlying clayey-silt (this thesis, Chapter 2). Samples were also taken from a thin (<4 mm) silty drape layer within the diamict of Whisler core #1, the 10 surface tills, and a soft sediment clast from the Appleby core #2 diamict. Chrysophyte stomatocysts were enumerated, but not described.

3.2.7 Chironomid Head Capsules

The chitinous head capsules of larval Chironomidae (non-biting midges) are often found well preserved in freshwater sediments and have a wide application to paleolimnological studies (Walker 1987). Species distribution and abundance are strongly correlated with summer surface-water temperature and maximum lake depth (as an index of heat storage capacity, Danks & Oliver 1972; Walker 1987; Walker *et al.* 1991). Head capsules were hand-picked from >63 μm sieved subsamples of the diamict, sorted sediments and overlying clayey-silt from Appleby, Brainard and Connell lakes.

3.3 RESULTS

3.3.1 Clast Fabrics

Stereo plots of the fabric data and summary statistics (cf. Mark 1973) are presented in Figure 3.5. For comparison, pebble fabric data from various glaciogenic sediments (Dowdeswell & Sharp 1986), including fossiliferous (ice-rafted) diamicts (Domack & Lawson 1985) and glaciogenic flow deposits (Lawson 1979), are also plotted (Fig. 3.6). Clast fabrics are also presented on a general shape triangle (Fig. 3.7) using indices of isotropy ($I=S_3/S_1$) and elongation ($E=1-(S_2/S_1)$)(Benn 1994).

The calculated Rbar values (Fig. 3.5) indicate that all of the fabrics are random at a 99 % confidence interval. Observations of random clast dip (*b*-axis) and plunge (*c*-axis), and

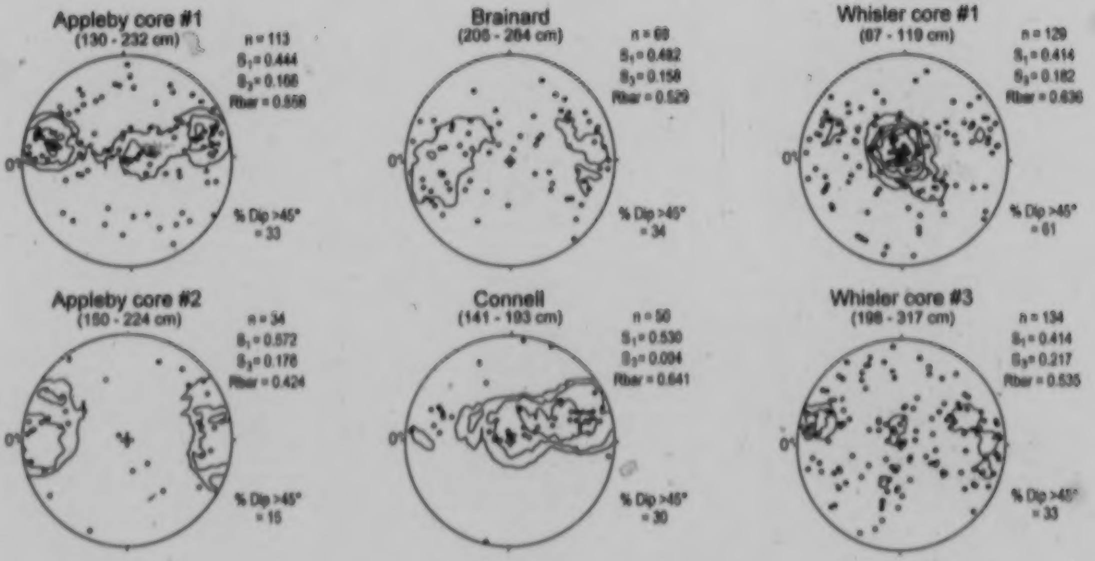

Figure 3.5 Schmidt equal-area stereo plots of clast fabrics from sections of the gravel-rich diamicts (GDmm, GDcm), contoured at 1 standard deviation. Depth of sample, clast numbers (n), eigenvalues (S_1, S_3), Rbar values and % of clast dip angles >45° are shown for each core sampled.

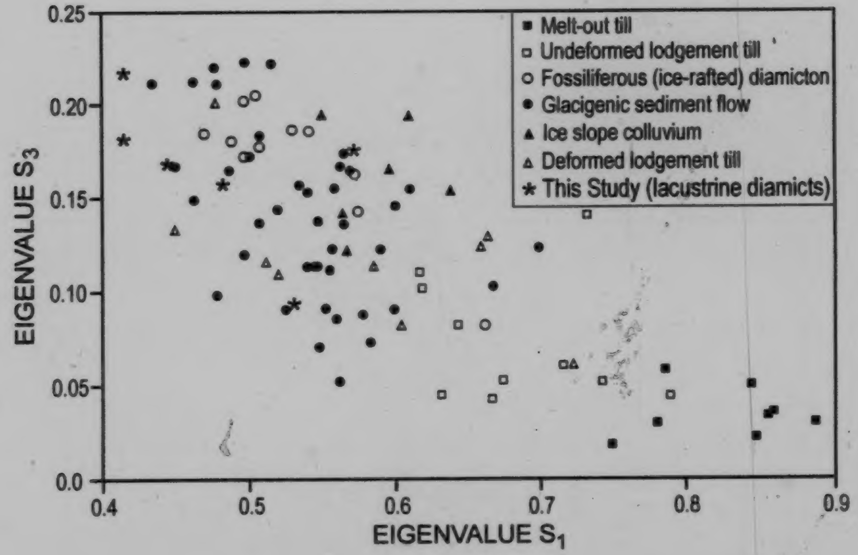

Figure 3.6 Eigenvalue S_1 vs. S_3 plot of the gravel-rich diamicts from this study compared to those of ice-rafted fossiliferous diamicts (Domack & Lawson 1985), glacigenic flow deposits (Lawson 1979) and tills (Dowdeswell & Sharp 1986).

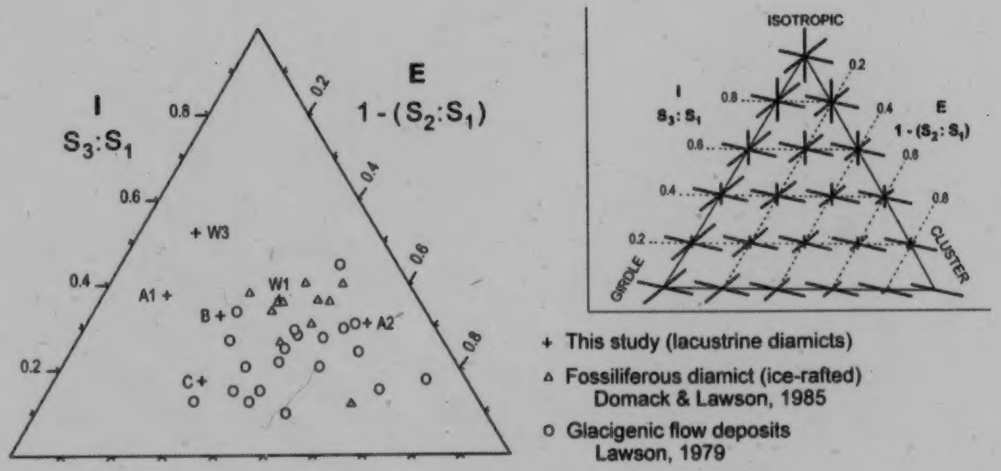

Figure 3.7 General shape triangle plot of the clast fabric eigenvalue data, using indices of Isotropy (I=S_3/S_1) and Elongation (E=1-S_2/S_1). Lacustrine diamicts plotted from this study are: A1-Appleby core #1; A2-Appleby core #2; B-Brainard; C-Connell; W1-Whisler core #1; W3-Whisler core #3. Inset shows the patterns of theoretical fabrics (*after* Benn 1994, Journal of Sedimentary Research, Figure 1)

the undisturbed nature of the diamicts and sorted sediment, would argue that the fabrics are not the product of core barrel warping, nor an artifact of the coring itself. Consider that the two Appleby cores were taken more than 100 m apart, with no record of either the core orientation to true north, or the position of the core tube when being split, yet still produce similar fabrics (Fig. 3.5).

On the general shape triangle (Fig. 3.7) the diamicts plot at a low to moderate isotropy (I=0.18-0.52) and elongation (E=0.11-0.56). The Whisler core #1 fabric was measured on what is an inversely graded, closed-work gravel (Fig. 3.8). A possible explanation for the relative strength of the central clustering (Fig. 3.5) is that this deposit may have formed by slumping, and thus reflects realignment, and clast to clast contact. Post-depositional settling and realignment of smaller clasts by larger ones may well have occurred in each of the diamicts. This would explain the single clast of coal within Connell Lake core, found vertically sheared into three separate clasts, encased within the finer matrix (2.3 cm vertical and 0.4 cm horizontal offset).

Less than 2 % of the clasts within the core diamicts were striated, whereas greater than 10 % of the clasts from the regional till samples were striated. Numerous clasts from the core diamicts had some secondary carbonate precipitate, while most from the surface till samples had some and often thick carbonate coatings. Several clasts from the diamicts of Whisler, Connell and Appleby lakes had filamentous algae attached to them, although none was found in the surface tills.

3.3.2 Particle-size

Gaussian component analysis revealed that 3 to 6 discrete modes occurred repeatedly within the diamicts and regional till samples, and that texturally, there was no discernible difference between them. This is not surprising given that the source of the diamicts, if not a till, could only be from the catchments, which themselves are covered by a till veneer. Clustering of the coarser sand and gravel fractions was similar among samples from each lake, but showed considerable differences between lakes. Sorted sediments from the Connell Lake core (120-141 cm) show a distinct coarsening, with little fine silt and clay compared to the underlying diamict (Fig. 3.9). These sediments were likely deposited in a

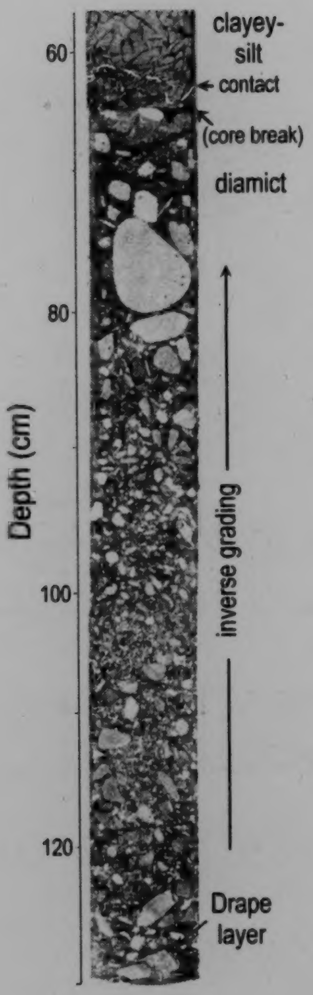
clayey-silt
contact
(core break)
diamict
inverse grading
Drape layer
Depth (cm)
60
80
100
120

Figure 3.8 Photograph of the lowermost section of Whisler core #1 showing the gravel-rich, clast supported diamict (GDcm) displaying inverse grading. Contact between the diamict and overlying clayey-silt (62.5 cm) is irregular, but sharp. The silty drape layer is found at 128.5 cm.

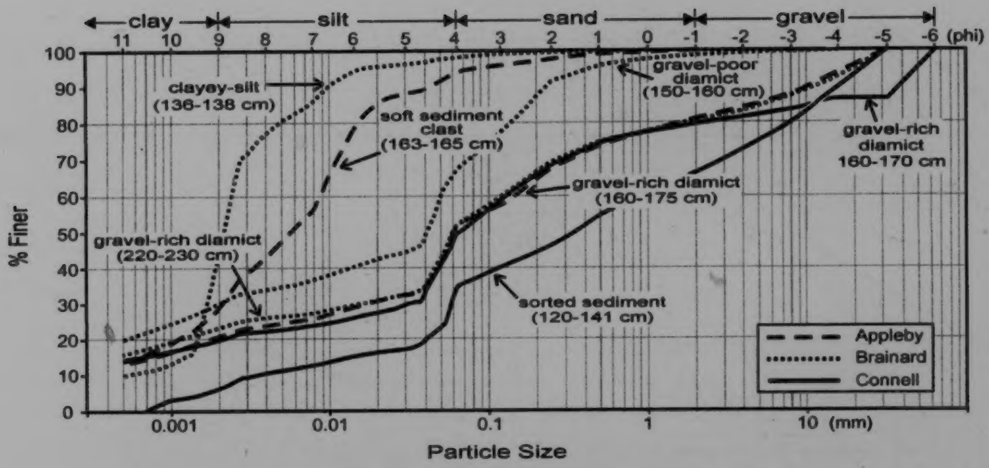

Figure 3.9 Particle size diagram highlighting the similarity of gravel-rich diamicts from three different lakes, and the marked differences between these and the more gravel-poor upper diamict of Brainard Lake. Current bedded sediments from Connell Lake display a distinct coarsening.

proglacial lake, impounded between the northward retreating ice and the plateau to the south. Fine silt and clay may have become decanted from the basin through conspicuous bedrock meltwater canyons. Sediments in the Brainard diamict show a fining upward trend through the upper, gravel-poor diamict into the overlying clayey-silt. The particle-size characteristics of a soft sediment clast from Appleby core #2 (163-165 cm), are unlike those of any of the diamicts, and are broadly similar to those of the overlying clayey-silt in Brainard Lake (Fig. 3.9). The origin of this and other soft sediment clasts found in both the Appleby Lake cores is uncertain, but may relate to 1) overriding and redeposition of proglacial lake sediments during glacial advance, 2) rafting of frozen sediments, or 3) slumping and/or lake ice-rafting of coherent blocks of lacustrine sediments which extend 1.5 m above the modern level of Appleby Lake, and encircle both Appleby and Rogers Lake.

3.3.3 Ostracodes

Ostracodes (Fig. 3.10, Tables 3.2 and 3.3) were found in Connell Lake at low abundances in the lower diamict and sorted sediments (180-195 cm), at trace levels within the middle diamictic sediments (141-180 cm), and at higher abundances in the upper diamict and sorted sediments (101-141cm). Ostracodes were abundant in the Brainard Lake clayey-silt, but found only in the upper 10 cm (150-160 cm) of the gravel-poor diamict. Within Appleby Lake core #1, ostracodes were found at low abundances in the upper 10 cm of the gravel-rich diamict (130-140 cm), were absent within the gravel-poor diamict between 130-121 cm, but were increasingly abundant through the upper gravel-poor diamict (121-100 cm) and overlying clayey-silt. Ostracodes were not found in any of the 10 regional till samples, including #50, taken from a moraine which impounds Connell Lake.

The ostracode assemblages are dominated by *Limnocythere liporeticulata* and to a lesser degree, *Candona rectangulata* (Table 3.3). There does not appear to be any significant difference between the assemblages in the diamicts and sorted sediment and those in the overlying clayey-silt (Table 3.3). Instar (juvenile) shells of all sizes were found within the samples, and are predominant in number (Table 3.2). Between 11 and 27 % of the ostracodes picked from all types of diamict remained articulated, while <5 % of the

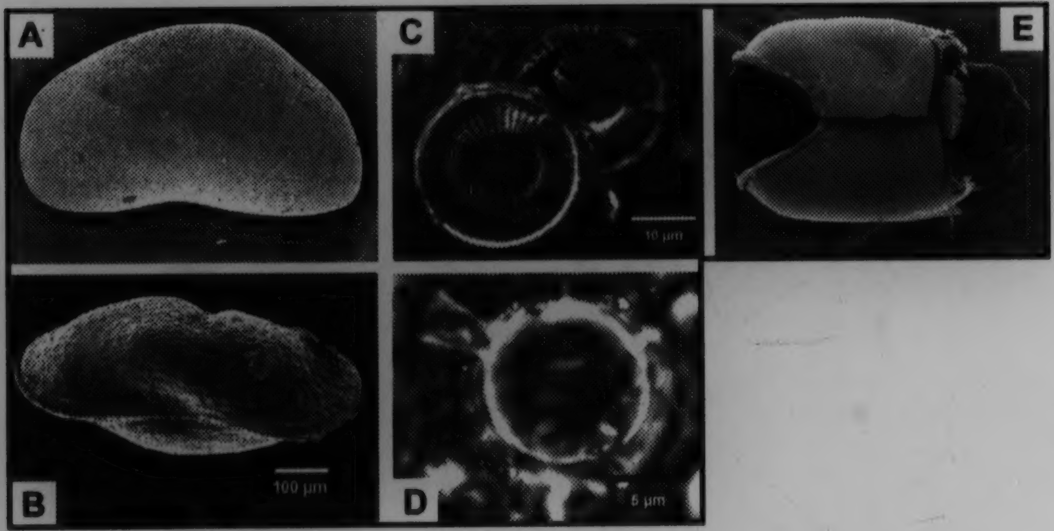

Figure 3.10 Photos and scanning electron micrographs of subfossil remains within lacustrine diamicts. **3.10A** ostracode, *Candona rectangulata*; **3.10B** ostracode (articulated), *Limnocythere liporeticulata*; **3.10C** diatom, *Stephanodiscus alpinus* Hust.; **3.10D** chrysophyte stomatocyst (species unknown); **3.10E** chironomid head capsule, ?*Tanytarsina* sp.

Table 3.2 Abundances of different biological subfossils from sections of the clayey-silt, massive, gravel-rich and gravel-poor diamict, and sorted sediments from cores of three lakes.

Sample Depth (cm)	Total dry sed. weight (g)	Dry sed. weight (g) < 2 mm	# Ostracodes	# Chironomid head capsules	# macro-fossil type a^2	Cladocera egg cases	Stratigraphy
			expressed as per gram dry sediment <2 mm[1]				
Appleby Lake core #1							
91-93	9.4	9.4	362.6	21.4	6.72	0.96	clayey-silt
99-97	9.7	9.7	291.5	16.3	4.39	1.41	clayey-silt
101-103	5.6	5.0	93.3	14.6	1.54	0.89	diamict, gravel-poor
111-113	11.8	10.0	21.0	3.1	0.26	0.13	diamict, gravel-poor
119-121	14.7	13.6	0.2	0.9	0.15	0.51	diamict, gravel-poor
123-125	19.2	15.1	0	0.3	0	0.20	diamict, gravel-poor
130-140	136.5	102.3	0.1	0.1	0.02	0.02	diamict, gravel-rich
Connell Lake core #1							
95-97	10.8	10.8	71.3	17.5	2.50	0.65	clayey-silt
97-99	9.7	9.7	301	25.5	3.30	0.72	clayey-silt
101-103	24.5	24.5	221.3	12.8	1.87	0.43	clayey-silt
103-105	53.8	21	32.8	4.9	0.48	0.81	diamict, gravel-poor
106-108	74.6	39	18.7	2.7	0.46	0.31	gravel-rich diamict + sorted sediment
112-114	68	41	41.9	5.0	0.68	0.12	diamict, gravel-rich
118-120	105.5	52.9	10.4	1.2			diamict, gravel-rich
120-141	141.6	93.8	0.6	0.3	0.04	0.01	sorted sediment
180-190	97.9	84.7	4.9	1.0	0.09	0.05	gravel-rich diamict + sorted sediment
190-195	68.2	52.2	5.0	1.8	0.19	0.08	diamict, gravel-rich
Brainard Lake core #1							
130-132	7.8	7.8	1.7	0.8	0.90	0.26	clayey-silt
140-142	16.8	16.8	170.5	57	7.02	1.31	clayey-silt
142-144	20.1	20.1	32.4	6.2	1.84	0.20	clayey-silt
148-150	22.9	22.9	4.3	2.4	0.22	0.04	clayey-silt
150-160	90.5	89.1	2.1	0.2	0.06	0.01	diamict, gravel-poor

[1] Sample concentrations are determined with respect to the dry sample weight less than 2 mm. This removes dilution effect by large clasts, as original sample sizes were insufficient to obtain a statistically representative sample of the > 2mm sized fraction (cf. Gale & Hoare 1992).

[2] Unidentified macrofossil: spherical, hollow, pale brown, approx. 0.4 mm diameter.

Table 3.3 Ostracode species composition (% of total) for clayey-silt and underlying massive and/or gravel-rich diamict from Connell and Brainard lakes cores.

Lake	Sample Depth (cm)	Total # Adult Ostracodes	Species % of total[1]					Stratigraphy
			Limnocythere liporeticulata	*Candona rectangulata*	*Prionocypris glacialis*	*Eucypris foveata*	*Limnocythere camera*	
Connell	0-3	49	79.6	20.4				clayey-silt
Connell	101-103	53	20.8	62.3	9.4	7.5		clayey-silt
Connell	180-190	85	63.5	36.5				diamict, gravel-rich
Connell	190-195	22	50	31.8	13.6		4.5	diamict, gravel-rich
Brainard	130-132	6	83.3	16.7				clayey-silt
Brainard	134-136	87	88.5	11.5				clayey-silt
Brainard	136-138	72	88.9	11.1				clayey-silt
Brainard	140-142	506	26.1	73.3	0.6			clayey-silt
Brainard	142-144	19	15.8	84.2				clayey-silt
Brainard	146-148	4	75	25				clayey-silt
Brainard	148-150	44	22.7	77.3				clayey-silt
Brainard	150-160	56	80.4	12.5	3.6	1.8	1.8	diamict, gravel-poor

[1] Ostracode species identified by D.L. Delorme

ostracodes in the purely sorted sediment samples (i.e., not a mix of diamict and sorted sediment interbeds) were articulated.

There is very little available autecological information on the species identified, other than that they record littoral, cold water environments (Delorme 1969; Røen 1981; Forester *et al.* 1989). *C. rectangulata* is common in stream environments and occurs occasionally in lakes north of 60° (Delorme 1973; Forester *et al.* 1989). It rarely occurs in lakes and ponds throughout the southern half of Greenland (Røen 1968, 1970), but was not reported by Røen (1981) in a survey of more than 60 freshwater environments (tarns, ponds, pool, streams) around Hazen Camp (Fig. 3.1, site 1). In the same survey, Røen (1981) reported that *Prionocypris glacialis* is generally found at low abundances but existed within 71% of the lakes, 50% of the streams, and 45% of the ponds. It has also been reported in lakes and ponds from northern Greenland (Røen 1962, 1968). Perhaps most interestingly, all five species are reported in Pleistocene sediments from the Yukon (Delorme 1968, 1969, 1974).

3.3.4 Diatoms and Chrysophyte Stomatocysts

More than 30 diatom species were identified in the stone scrapes and diamict matrices (Appendix 3.1, Fig. 3.10). There were also numerous fragments, which could only be identified at the genus level. Chrysophyte stomatocysts were present, often in comparable abundance to diatoms (Appendix 3.1, Fig. 3.10). No diatoms or chrysophyte stomatocysts were detected in the soft sediment clast or surface tills. Sampling techniques did not allow a determination of relative diatom densities, but concentrations were visually similar to sections of the overlying lacustrine sediment interpreted as having been deposited during periods of extensive summer ice cover ($<10^5$ valves/gram dry sediment). The diatom assemblage and percentage of different species from the drape layer in Whisler core #1 (Fig. 3.8) is quite similar to the overlying lacustrine sediment, interpreted as having been deposited during a period of extensive summer ice cover (Fig. 3.11). This suggests that the drape layer was deposited within a lake environment with an *in situ* benthic population.

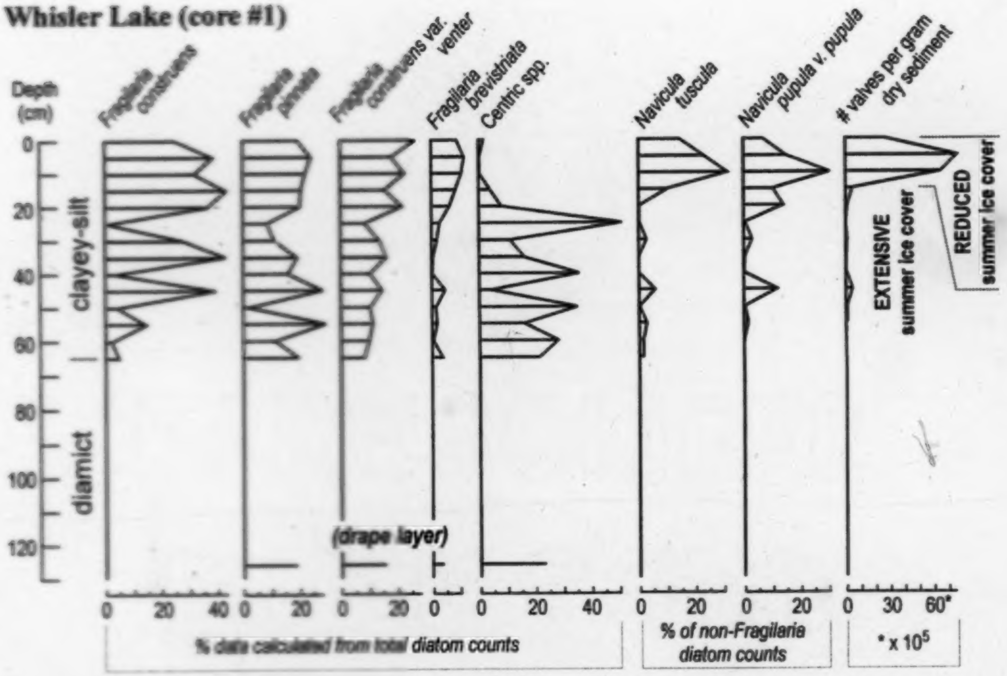

Figure 3.11 Diatom % abundance for Whisler core #1, highlighting the species compositional similarity between the drape layer (128.5 cm) and much of the early post-diamict clayey-silt, interpreted to have been deposited during a period of extensive summer lake ice cover.

3.3.5 Chironomid Head Capsules

The chironomid assemblages from six samples of the diamicts, sorted sediment and clayey-silt from Brainard and Connell lakes are markedly similar (Table 3.4). Low concentrations (≤2/g dry sediment) of head capsules (Fig. 3.10) were found in all samples of the core diamicts and sorted sediments (including samples devoid of ostracode remains) from Appleby, Brainard and Connell lakes (material from Whisler Lake was not scanned). Concentrations from the immediately overlying clayey-silt were one to two orders of magnitude greater, and displayed similar abundances as the ostracodes (Table 3.2). The species identified are consistent with very cold water, many of which (*Heterotrissocladius*, *Sergentia*, *Stictochironomus*, *Oliveridia* and especially *Paracladius* and *Pseudodiamesa*) are considered to be cold-stenotherms (Walker, pers. comm. 1996). Although very abundant, *Tanytarsina* is always so, and therefore essentially meaningless in terms of autecological considerations (Walker, pers. comm. 1996).

3.4 DISCUSSION

3.4.1 Deglacial History and Diamict Chronologies

At the LGM all lake basins investigated in this study were overridden by ice that emanated from the Grant Land Mountains and Agassiz Ice Cap, covering Hazen Plateau, and was coalescent with the Greenland ice margin to the east (England, in press). Extensive field mapping and dating of raised marine deposits indicate that by ~7.5 - 7 ka BP, glaciers along the margin had thinned considerably (<300 m asl along the eastern Lake Hazen Basin), and contacted the sea at the head of Chandler Fiord and Black Rock Vale (Fig. 3.12). Radiocarbon dates of 7860±270 BP (GSC-3179; England 1983) and 7550±90 BP (TO-4474; Table 3.5) on shells from marine silts underlying gravel terraces provide minimum ages for the estimated 87 m marine limit at the head of Chandler Fiord. The ice margins forming these marine limit terraces, their subsequent pattern of retreat, and the corresponding ice configuration across Lake Hazen Basin, are recorded by *en echelon* meltwater channels, moraines, kames and kame deltas (Fig. 3.12). Small, arcuate moraines, lateral and proglacial meltwater channels, and other geomorphic evidence indicate that the upper plateau (including Brainard and Hart lakes) was covered by local plateau ice caps, subsequent to the

Table 3.4 Abundance and species composition (% of total) of chironomid head capsules for differing sediment types within the Connell and Brainard cores.

Chironomid sp.[1]	Chironomid Head Capsules (species % of total)					
	Connell L. 3-5 cm (clayey-silt)	Connell L. 103-105 cm (diamict, gravel-poor)	Connell L. 106-108 cm (diamict + sorted sed.)	Connell L. 118-120 cm (diamict, gravel-rich)	Brainard L. 134-136 cm (clayey-silt)	Brainard L. 150-160 cm (diamct, gravel-poor)
subtribe Tanytarsina	54.1	54.3	62.1	69.0	67.0	66.7
tribe Pentaneurini	1.5	16.3	1.6	4.0	12.3	
Heterotrissocladius	17.7	9.2	11.3	14.0	2.8	
Paracladius	17.1					
Procladius	3.6	4.3	10.5	4.0	13.2	33.3
Sergentia	3.0	6.5	5.7	3.0		
Stictochironomus	0.4	3.3	1.6		1.9	
Psectrocladius	0.3					
Oliveridia?	0.3					
Cricotopus/Orthocla	0.1		4.0	1.0		
Chironomus		1.1			1.9	
Pseudodiamesa					0.9	
unidentifiable	1.8	4.9	3.2	5.0	0.0	0.0
Total # Chironomid Head Capsules	335.5	92.0	62.0	50.0	53.0	3.0

[1] Species identified by I.R. Walker.

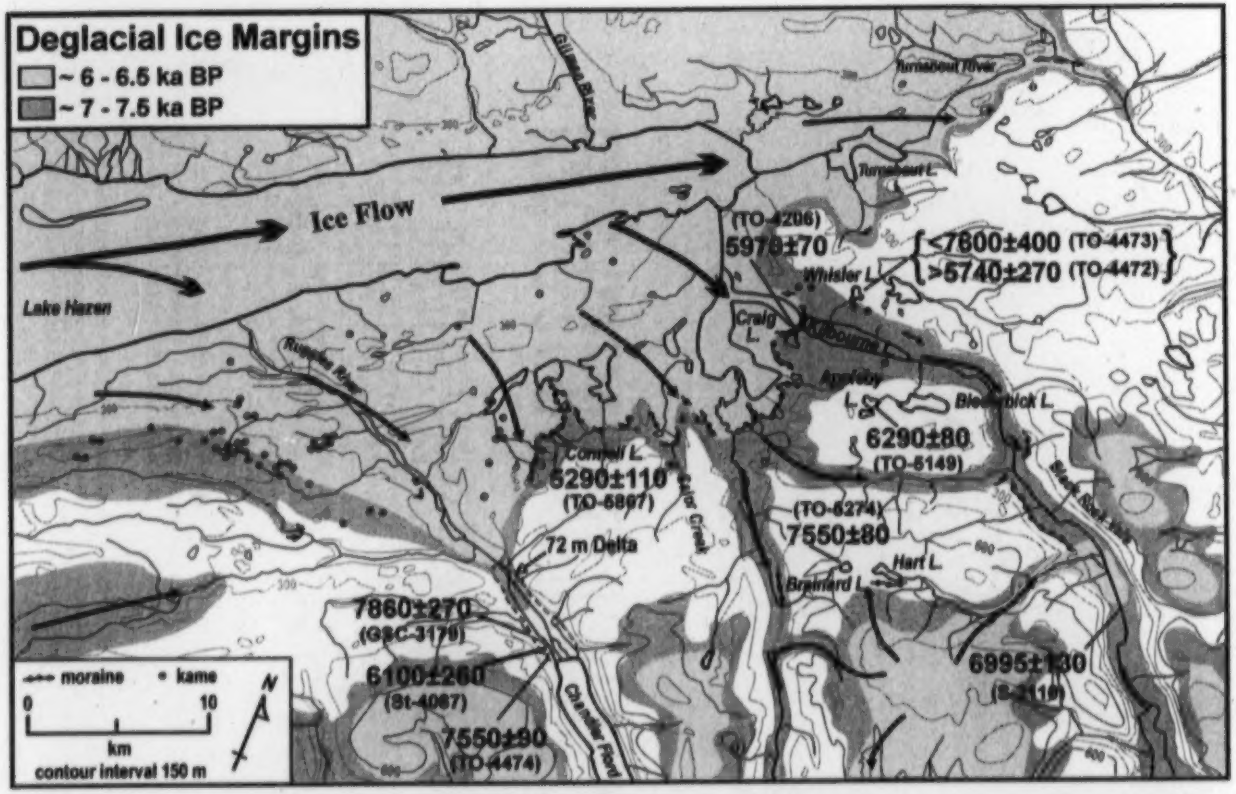
Deglacial Ice Margins
~ 6 - 6.5 ka BP
~ 7 - 7.5 ka BP
Ice Flow
Lake Hazen
Gilman River
Turnabout River
Turnabout L.
Ruggles River
(TO-4206)
5970±70
Whisler L.
<7600±400 (TO-4473)
>5740±270 (TO-4472)
Craig L.
Kilbourne L.
Appleby L.
6290±80
(TO-5149)
Connell L.
6290±110
(TO-6897)
Salor Creek
(TO-6274)
7550±80
Hart L.
Brainard L.
Black Rock Vale
72 m Delta
7860±270
(GSC-3179)
6100±260
(St-4087)
7550±90
(TO-4474)
Chandler Fiord
6995±130
(S-2119)
300
600
moraine
kame
0
10
km
contour interval 150 m
N

Figure 3.12 Deglacial ice margins for 7.5 - 7 and 6.5 - 6 ka BP. Margin reconstructions are based on radiocarbon dating of raised marine deposits, and interpretation of the regional glacial geomorphology (deltas, moraines, kames, meltwater channels and striae). The region was dominated by a large trunk glacier advancing northeastward up the Lake Hazen trough, which spilled obliquely across the lowlands and valleys south of Lake Hazen. Ice margins drawn on the upper plateau (along the bottom of the diagram) are for the most part poorly constrained, and best considered provisional.

Table 3.5 Radiocarbon date list and sample data. All dates are reported with one standard deviation.

Site	Laboratory Dating No.	Material[a]	Age[b]	Stratigraphy	Core Depth (cm) / Elevation (m asl)
Appleby core #2	TO-5148	*Drepanocladus fluitans*	4220±60	clayey-silt	70 cm
Appleby core #2	TO-5149	*Scorpidium turgescens*	6290±80	clayey-silt	102 cm
Brainard core #1	TO-5272	*Drepanocladus aduncus var. kneiffii*	3030±110	clayey-silt	29 - 31 cm
Brainard core #1	TO-5274	*Scorpidium turgescens, S. scorpioides*	7550±80	clayey-silt	141 - 144 cm
Connell core #1	TO-5867	*Drepanocladus aduncus var. kneiffii*	5290±110	sorted sediment	168-170 cm
Whisler core #1	TO-4473	*Drepanocladus brevifolius*	7600±400	diamict, gravel-rich	80.5 cm
Whisler core#1	TO-4472	*Scorpidium turgescens*	5740±270	clayey-silt	27 - 28 cm
Craig Lake Moraine	TO-4206	peat	5970±70	*in situ* peat underlying ice contact delta	176 m asl
Chandler Fiord	TO-4474	*Portlandia arctica*	7550±90	marine silts underlying 87 m gravel terrace	58 m asl

[a] Moss species identified by C. LaFarge-England

[b] Moss and peat samples were subjected to an Acid-Alkali (AAA) pretreatment by IsoTrace Laboratories, designed to remove humic and tannic contaminants and isolate the lignin and cellulose fractions. Shells were subjected to an acid leach (Beukens, pers. comm. 1996).

retreat of the trunk glacier towards the Grant Land Mountains. Deglaciation of Brainard and Hart lakes occurred when ice retreated southwards onto the plateau, and is constrained by the minimum date of 7550±80 BP (TO-5274, Table 3.5; Fig 3.12) from Brainard Lake, measured on aquatic moss from sediment (141-144 cm depth) overlying the diamict (145 cm).

Below the 87 m marine limit, an extensive 72 m delta extends for several kilometres up the Ruggles River (Fig. 3.12). Shell samples from the foreset beds date 6100±130 BP (St 4087; England 1978) and 6255±110 BP (S-1980; England 1983). A discontinuous moraine extends from the 72 m delta, northeastward to the Craig Lake Moraine (Fig.3.12). The Craig Lake Moraine is constrained by a minimum date of 5970±70 BP (Table 3.5) on peat buried by an ice-contact delta emanating from the moraine. Bender and Connell lakes are impounded by sections of this morainic belt (Fig. 3.12). Deglaciation of Connell Lake is constrained by a date of 5290±110 BP (TO-5867; Table 3.5) determined on mosses from a thin lens (172-174 cm depth) of sorted sediment (containing *Oxyria digyna* seeds, balls of *Nostoc* algae, and the coalified stem and roots of *Salix arctica*) in between two sections of diamict (Fig. 3.3). If the Connell Lake diamicts are tills, then this date provides a maximum age for deglaciation of the basin. However, if the diamicts were formed pro- or, extra-glacially, than the date provides a minimum age for deglaciation.

The striae and geomorphic records of the trunk glacier occupying Lake Hazen Basin indicate that the Biederbick/Appleby Lake upland is glacially downflow from the Chandler Fiord marine limit site (Fig. 3.12); therefore, 7860±270 BP represents a minimum age for deglaciation of these basins (elevations of 440 and 447 m, respectively; Table 3.1). As there is no stratigraphic or geomorphic evidence of a subsequently formed plateau ice cap over this upland, the radiocarbon date of 6290±80 BP (TO-5149; Table 3.5) on aquatic moss from the sediments 1 cm above the diamict in Appleby Lake clearly represents a minimum age for deglaciation.

During subsequent deglaciation, ice remained in the Kilbourne Lake valley, having retreated from the Biederbick/Appleby Lake upland, and from the opposite side of the valley, where Whisler Lake is situated (Fig. 3.12). This retreat is well documented by the regional geomorphology, in particular the staircase of deltas deposited in extensive proglacial lakes. The upper diamict contact (62.5 cm depth) in Whisler core #1 is bracketed by radiocarbon dates of 7600±400 BP (TO-4473; 80.5 cm depth) within the diamict, and 5740±270 BP (TO-4472; 27-28 cm depth) within the clayey-silt (Fig. 3.3; Table 3.5). The large standard errors in both samples may be explained by the small amount of material submitted for dating; a single moss fragment comprised each sample. The 7.6 ka BP date must be treated cautiously as the moss' origin is clearly allochthonous, and its presence within an inversely graded deposit (slumped?) suggests further reworking. Nevertheless, the rate of sedimentation in Appleby and Brainard lakes (Fig. 3.13) would argue that 2000 years for the 34 cm of clayey-silt accumulated between the upper diamict contact and the 5.7 ka BP sample is reasonable. The 7.6 ka BP date thus provides a maximum age for the upper diamict contact, and a minimum age for deglaciation of the basin. If the diamicts are in some manner related to the local retreat of glaciers then the date overlying the Appleby Lake till of 6.3 ka BP is clearly too young.

3.4.2 Tills, Flow or Ice-Rafted Deposits?

While the chronology of the diamicts from Brainard and Whisler lakes appear to coincide with the pattern of deglaciation, it is not proof that the diamicts are tills. In fact, the clast fabric data (Fig. 3.5) clearly plot within the range of fossiliferous (ice-rafted) diamict facies and glaciogenic flow deposits (Figs. 3.6 & 3.7; Lawson 1979; Domack & Lawson 1985; Dowdeswell & Sharp 1986), clearly distinguishing them from those of basal ice, melt-out and lodgement till. This distinction is further supported by the presence, and extreme fragility of many of the subfossils, especially the ostracodes (Westgate & Delorme 1988) and chironomid head capsules, which could not survive glacial abrasion or granulation. In rare cases, however, preservation and reworking of soft-sediment clasts and subfossils such as ostracodes can occur where there is brittle shear of nondilatant, coherent blocks of sediment, and ductile deformation of dilatant sediment, as in deformation tills (cf. Hicock & Dreimanis 1992; Benn & Evans 1996). However, in the deformation till fabrics of Hicock & Dreimanis (1992), there is not the high percentage of dip angles >45° seen in this study (Fig. 3.5).

The clast fabric from a diamict section of Connell Lake (Fig. 3.5) also disputes a till genesis for the second diamict facies assemblage (alternating diamict and sorted sediment). The sedimentology of diamicts from this second facies further supports this

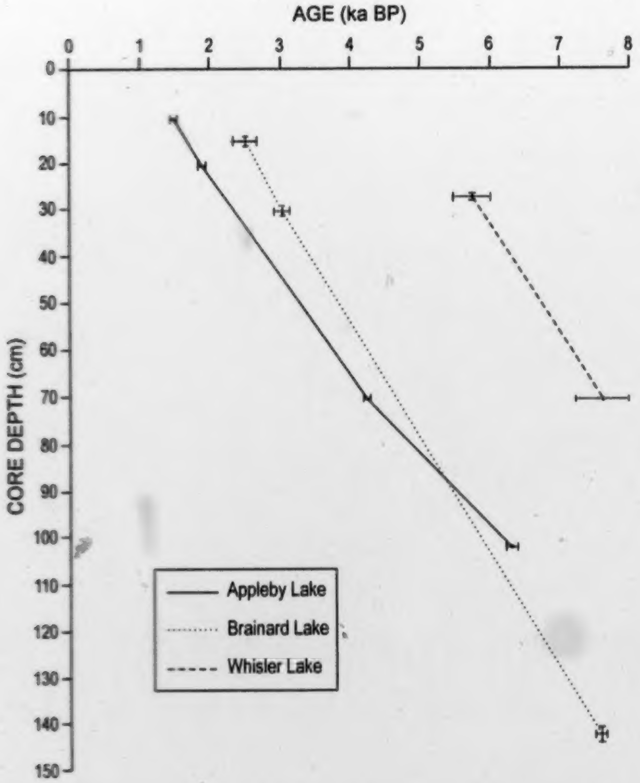

Figure 3.13 Age vs. depth relationships. Error bars for core depths correspond to the depth of material from which the organic material was extracted (single fragments are indicated by a point), while for radiocarbon dates, error bars represent one standard deviation. Figure demonstrates very similar sedimentation rates for each of the three lakes. Note, approximately 50 cm of the uppermost sediment of Whisler Lake core #1 was lost in the field, accounting for its displacement from the other two lake records.

interpretation, as lower contacts of diamict sections are conformable to bedding structures below.

Glaciogenic flow deposits have a tendency to more pronounced elongation and clustering (Lawson 1979; Dowdeswell & Sharp 1986; Figs. 3.6 and 3.7). As discussed by Lawson (1979) this principally reflects the water content of the flows; those with higher water contents (18-25 % water by weight) have more consistent clast alignment. The random clast fabric, high percentage of dip angles >45° (15-61%; Fig. 3.5), and low elongation values of diamicts from this study (Fig. 3.7), argues against their being flow deposits, and instead favours an origin as a rain-out deposit. The absence of sorting within the diamicts also argues against a flow hypothesis.

3.4.3 Ice-Rafting

Glacial rafting (including iceberg dump and rainout, and deposition from an iceshelf) and sea ice rafting (by adfreezing, frazil and anchor ice) have long been recognized as important transport and depositional processes in the marine environment, and are linked to the formation of gravel-rich, massive diamicts (Barnes *et al.* 1982; Gilbert 1990a; Anderson *et al.* 1991; Dowdeswell *et al.* 1994). Similar processes have been inferred and shown to occur in proglacial lakes (Eyles *et al.* 1983; Gilbert & Desloges 1987; Rovey & Borucki 1995), as well as modern lakes, such as Lake Michigan (Miner & Powell 1991; Barnes *et al.* 1994).

A number of studies of Arctic lake sediments have suggested, or provided evidence, for ice-rafting as the origin of isolated clasts or lenses of coarse-grained sediments (sand and gravel; Røen 1962; Coakley & Rust 1968; Shilts & Dean 1975; Ugolini 1975; Vorren 1978; Gilbert & Church 1983; Fredskild 1995). Blake *et al.* (1992) studying lake sediments from Kap Inglefield Sø, northwest Greenland (Fig. 3.1, site 2), illustrate a core containing gravel and discrete sand lenses throughout the uppermost 49 cm and attribute these to rafting of littoral sediments by the lake ice. Nichols (1967; 1974) reported on a 3-4 m deep, "mud-centred" lake in which massive blocks of sediment-rich ice, several metres in length and at least one metre thick, rose to the surface a week after the surface cleared of ice. These blocks melted over several days, during which sediment was deposited randomly across the basin by rainout and iceberg dump (when the blocks became unstable and rolled over).

Further support for the interpretation that the diamicts formed by ice rafting (as opposed to the deposition of glacial debris) is that the clasts and fines appear to be reworked littoral lacustrine sediment. In Appleby, Brainard and Whisler lakes, the clayey-silt is dominated (63-97%) by benthic *Fragilaria* spp. (this thesis Chapter 2). Yet, within the diamict samples from these lakes, the diatom species are almost exclusively littoral and planktonic forms (Appendix 3.1). Planktonic species are extremely rare in the clayey-silt of these lakes, and never occur in the absence of some benthic population, as they do in the diamicts. Similarly, within the clayey-silt, littoral diatoms are very rarely found, and never in the abundance, or diversity as seen in the diamicts. The unusual diatom assemblages within the diamicts, clearly highlight the uniqueness of the process responsible for deposition of the diamicts. Scans of the surface till samples indicate no presence of diatoms or any of the other subfossils investigated (ostracodes, chironomid head capsules, filamentous algae).

While the diamicts do not appear to be formed from the direct deposition of glacial debris, the abruptness with which they end, the general correlation of their chronology to the regional retreat of glaciers, and the absence of similar ice-rafted debris elsewhere through the Holocene sedimentary records, clearly argues that they are somehow linked to the presence of glaciers.

All of the lakes displaying the primary diamict facies (Appleby, Beiderbick, Brainard, Stewart and Whisler lakes) are small, topographically isolated basins. They also all record evidence of higher lake levels, that could only have existed when glaciers impounded the regional drainage. In the Whisler Lake basin, ice retreating into the Kilbourne valley deposited two kame deltas 21 and 10 m above the northwestern shore of Whisler Lake, for which there are accordant outlet channels on the eastern shore draining into the adjacent valley (Fig. 3.14). Similarly, a moraine and small raised ice-contact delta behind Hart Lake mark the southward retreat of ice. In both lakes, the deltas are composed of glacial drift, in which fines and coal are abundant. Despite this, there is no proximal-distal fining, or any evidence of sorting, observed within the four cores from each of these two small basins. Sorted sediments also don't exist in the diamicts of the other three lakes belonging to this facies group. Given that in all other basins, which can be linked to

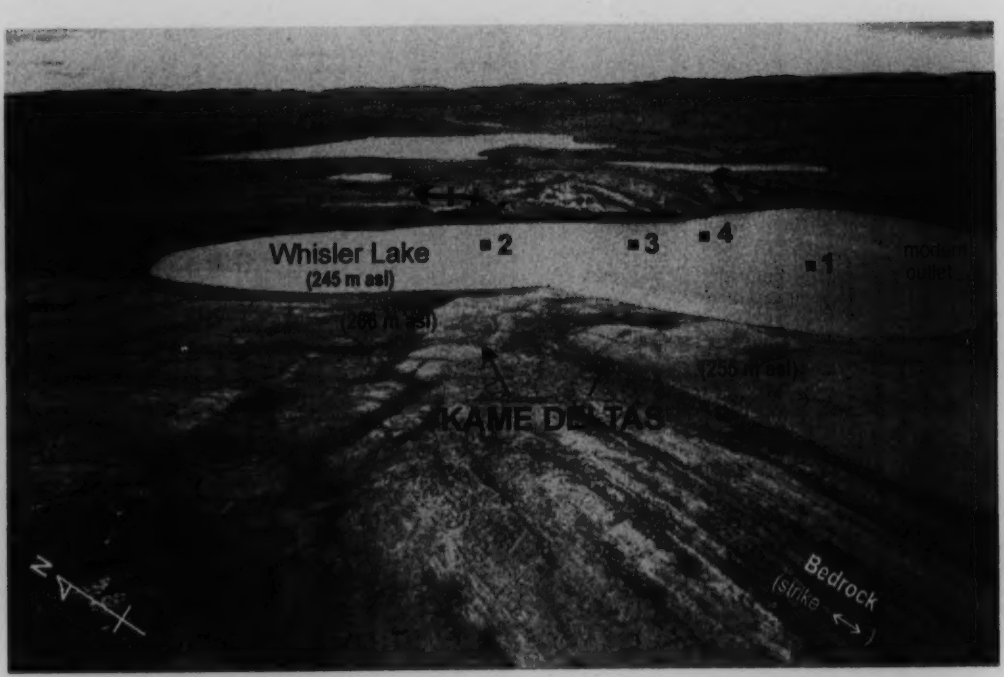
Whisler Lake
(245 m asl)
■ 2
■ 3
■ 4
■ 1
modern outlet
(266 m asl)
(256 m asl)
KAME DELTAS
Bedrock
(strike ↔)
N

Figure 3.14 Oblique photograph of the Whisler Lake basin showing the two kame deltas deposited when retreating ice impounded the modern drainage into Kilbourne Lake, and instead directed flow through conspicuous outlet channels (hachured arrows) into the adjacent valley.

proglacial sedimentation (Bender, Connell, Gardiner, Hodges, Kilbourne, Linn) there is a clear and abundant stratigraphic record of proglacial deposition (Fig. 3.3), then it is odd that there would be no depositional record within Whisler and Hart lakes relating to the emplacement of the deltas. The absence of such in the cores from Whisler and Hart lakes, suggests that it may exist below the recovered diamicts, and hence the diamict postdates the deposition of these features. However, as these ice contact features are raised above the modern bedrock outlets of each of these basins (1.5 - 10 m), then there was likely an undefined period of time of higher lake levels, during which there is no direct evidence for deposition of sorted sediment or ice-contact features. The higher lake levels still require impoundment by glaciers, but it may be that sediment-laden meltwater was simply diverted elsewhere, or that sediment-traps existed between the glacier margin and the proto-lake basin. There may well have been a point at which the lakes were glacier-fed, but not directly proglacial.

3.4.4 A Possible Mechanism For Ice-Rafting

In lakes experiencing complete seasonal ice cover, layers of adfrozen littoral sediment (including larger boulders) may be plucked and ice-rafted small distances as a result of the sudden rise in spring lake levels brought about by the ponding of meltwater by slush accumulations at the outlet (Shilts & Dean 1975; Gilbert 1990b). While this same mechanism of damming of the outlet and buoyant lifting of the ice floe occurs on High Arctic lakes today (Woo 1980), the absence of ice-rafted debris throughout much of the Holocene benthic sedimentary records in this study argues that adfrozen sediment is released by basal melting and marginal thinning, prior to any significant breakup and mobilization of the ice floe.

In the case of a proglacial or glacier-fed lake, several factors may contribute to more extensive rafting of littoral sediments. One is that the range in water levels of the lake during the spring melt is likely to be consistently greater than from the nival recharge seen today. This would not only ensure a buoyant lifting of the ice pan, and possible plucking of adfrozen sediment, but would also act to more effectively break up the ice pan itself. This is critical to redepositing adfrozen sediment from the littoral zones to the deeper basins.

Large variations in water level associated with glacier-dammed lakes would also contribute to the reworking of raised lacustrine sediments. The second factor relates to the proximity of glaciers and inherent microclimatic effects. Summer temperatures were likely cooler in regions immediately adjacent the retreating glaciers, and katabatic winds would have more effectively scoured snow from lake ice surfaces. Both of these factors would have contributed to greater depths of freezing within the lakes. Conditions may have mimicked what is seen in modern-day Antarctica lakes, where ice thicknesses on the order of 2.8 to 6 m are common (Wharton *et al.* 1993). At Lake Hoare, Antarctica (ice thickness >3 m) the mean annual air temperature is -17.3°C (Clow *et al.* 1988), while at Lake Hazen (ice thicknesses ~2 m) it is -20°C (Thompson 1994). This demonstrates the significance of wind scouring of snow (Antarctica), and how the absence or diminishment of the summer "oasis" conditions in Lake Hazen Basin would have created a colder climate, leading to greater lake ice thicknesses (cf. Doran *et al.* 1996).

If the thickness of the ice floe is hypothetically extended from ~2 m today, to 3 m, there is an accordant 13 to 26% increase in lake bottom area which would become seasonally frozen onto the ice floe (Fig. 3.15). If Whisler and Hart lakes froze to a depth of 4 m, 69 and 73% of the respective basins would freeze to the lake ice. Not only does this increase the area frozen to the bed, but extends the range of freezing 10's to 100's of metres out from the shore (Figs. 3.2 and 3.15). Shallower lakes, such as Brainard may have become completely frozen. Thus, glacial meltwater input in spring would lead to a rapid rise in lake level, leading to extensive plucking of adfrozen lake sediment, and the potential rafting of littoral sediments across the basin.

The ostracode and chironomid assemblages show no differences between the diamict and overlying clayey-silt (Tables 3.3 and 3.4). Variability in the subfossil record, specifically actual numbers of subfossils found (Table 3.2), likely has as much to do with uncertainties in the sedimentation rate, as with environmental changes. Yet the fact that the same species are found both within the diamict and clayey-silt argues that some aspects of the lake's ecology must have remained similar. Both ostracodes and chironomids require minimum temperatures to be achieved in order for them to mature and progress through their larval or moult stages. Temperate and low Arctic lake studies suggest these may be around

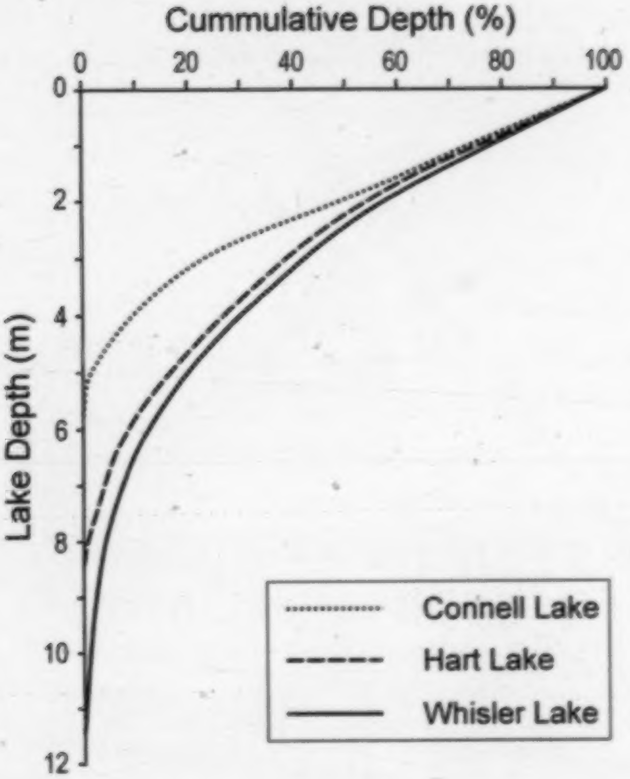

Figure 3.15 Graph shows cumulative depth (%), highlighting the large extent of each basin that exists above 4 m depth. Changes in the depth of freezing from ~2 m today, to >3 m hypothesized during deglacial times, leads to a considerable change in the extent of lake bottom seasonally frozen to the ice pan.

6-10°C for the cold-tolerant species seen in this study (Walker 1987; Delorme 1991), however, this remains to be documented in High Arctic lakes. Doran *et al.* (1996) indicated that the degree to which Antarctic lakes contact the glacier margin may determine whether they maintain a thick perennial ice cover. Thus, it is unlikely that the lakes in this study could be extensively glacier-contact, and yet achieve lake water temperatures high enough to support the growth and reproduction of ostracodes and chironomids. Therefore, the suggestion that the diamicts formed during final retreat of glaciers from the basin, or at a time when the lakes were only glacier-fed seems most plausible.

Other processes which may have contributed to the movement of debris are aeolian (katabatic winds) redistribution of raised lacustrine sediments onto the surface of the ice pan. Previous Arctic studies have demonstrated the capacity for winds to move cobbles from gravel sandurs well out onto fiord ice (McKenna-Neuman & Gilbert 1986). Yet the absence for relict dunes or deflation pavements surrounding these lakes (which do exist elsewhere in the study area) questions the relative importance of this. Increased wind action, especially during fall freeze-up may also have led to extensive anchor ice formation and entrainment of debris. However, the movement of the debris towards the centre of the lake would still be inhibited without some more effective means of breaking up the lake ice cover.

3.5 CONCLUSIONS

Detailed analysis of clast fabrics and subfossil remains supports the interpretation that the diamicts in cores of lake basins from the Lake Hazen region are not tills. Diamict clast fabrics are broadly similar to those measured by Domack & Lawson (1985) on fossiliferous diamicton interpreted as having formed by glacial ice rafting. They are also quite similar to some glaciogenic debris flows (Domack & Lawson 1985; Dowdeswell & Sharp 1986). However, the diamicts from this study are distinguished from glaciogenic debris flow deposits by the high frequency of clast dip angles >45° (15-61%). The absence of sorted sediment within the diamicts also argues against a debris flow origin.

The sedimentological records further dispute a proglacial origin for the diamicts. Abundant sorted sediments found in the cores of several of the lake basins, clearly record deposition in extensive proglacial lakes. Yet even where there is evidence of proglacial sedimentation in the basins containing the diamict facies (kame deltas beside Whisler Lake, and a raised delta in Hart Lake), there is no sedimentological evidence related to these depositional features within multiple cores from each of these basins. Thus, the diamict is considered to postdate this, and sorted sediment and possibly till, if present, would exist below the depth of recovered sediment.

The supposition that the diamict formed in a lake environment is supported by the presence of diatoms, ostracodes, chironomid head capsules, and plant detritus. Species assemblages of the ostracodes and chironomids show no major difference between the diamict and immediately overlying clayey-silt. This suggests that lake water attained sufficient summer warmth to allow full maturation of these species. Hence, the lakes are unlikely to have been extensively glacier-contact as this would lead to a perennial ice cover. The diatom assemblages from clast scrapes and diamict matrices, particularly the dominance of littoral and planktonic species and the low diversity and numbers of benthic species, supports a littoral, lacustrine origin for the sediments comprising the diamicts.

The near absence of clasts throughout the overlying clayey-silt, the abruptness with which the deposition of the diamict stopped, and the relative correlation of the diamict chronology with that of the regional glacial retreat strongly suggest some manner of glacial control. Thus a scenario in which the retreating glaciers either continued to impound part of the regional drainage within the basins, or at least fed meltwater into the lakes is proposed. The proximity of glaciers to these basins combined with the effects of katabatic winds would ensure that lake ice thicknesses would have been greater, allowing the depth of seasonal freeze-on to extend 10's to 100's of metres out from the shore. The mechanism for ice rafting of the littoral sediments then becomes the buoyant lifting of the ice cover, plucking of adfrozen sediment, and the rapid break up of the ice floe by continuing glacial meltwater inputs.

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3.7 APPENDIX

Appendix 3.1

Diatom and chrysophyte stomatocyst remains from diamict stone scrapes and matrix samples. Diatom species informally classified as "littoral forms" are those found extremely rarely, or absent, in benthic sediment cores, and relatively abundant in samples of the various littoral communities (epipelon, epiphyton, epilithon) in a survey of regional lakes and ponds (Smith, unpublished data). Undifferentiated forms included species that commonly occur in both the lake sediment cores and littoral samples.

Appleby Lake, core #1, 230-236 cm, stone scrape

Planktonic forms	
Aulocoseira subarctica (O.Müller)Haworth	3
Cyclotella pseudostelligera Hust.	1
Littoral forms	
Caloneis silicula (Ehr.)Cleve	4
Cymbella naviculiformis Auerswald	1
Eunotia circumborealis L.-B. & Nörpel	2
Eunotia praerupta Ehr.	1
Gyrosigma sp. (fragment)	1
Navicula incerta Lange-Bertalot	2
Neidium sp. (fragment)	1
Stauroneis phoenicenteron (Nitzsch)Ehr.	2
Stauroneis sp. (fragment)	1
Tabellaria flocculosa (Roth)Kütz	4
Undifferentiated forms	
Achnanthes minutissima Kütz.	6
Fragilaria construens var. *venter* (Ehr.)Grun	7
Nitzschia sp. (fragment)	1
Chrysophyte stomatocysts	27

Appleby Lake, core #1, 163-165 cm, soft sediment clast

no diatoms or stomatocysts seen

Appleby Lake, core #2, 220-224 cm, stone scrape

Planktonic forms	
Aulocoseira sp.	2
centric sp. (fragment)	1
Littoral	
Diploneis sp. (fragment)	1
Pinnularia sp. (fragment)	1
Tabellaria sp. (fragment)	2
Undifferentiated forms	
Achnanthes minutissima Kütz.	2
Navicula sp. (fragment)	2
Chrysophyte stomatocyst	2

Appleby Lake, core #2, 150-160 cm, stone scrape

Planktonic forms	
Aulocoseira distans	1
Chrysophyte stomatocysts	5

Brainard Lake, 260-267 cm, stone scrape

Planktonic forms	
Aulocoseira subarctica (O.Müller)Haworth	1
Aulocoseira sp.	5
Cyclotella sp.	2
Stephanodiscus cf. *minutulus* (Kütz)Cleve & Möller	3
Stephanodiscus niagarae Ehr.	2
Littoral forms	
Pinnularia sp.	2
Undifferentiated forms	
Fragilaria pinnata Ehr.	2
Achnanthes sp.	1
Chrysophyte stomatocysts	1

Brainard Lake, 220-230 cm, stone scrape

Planktonic forms	
Aulocoseira distans (Ehr.)Simonsen	2
centric sp. (unknown)	5
Littoral forms	
Eunotia sp. (fragment)	1
Fragilaria hungarica var. *tumida* Cleve-Euler	5
Surirella turgida W.Smith fo. *obtusa* Foged	10
Chrysophyte stomatocysts	20

Brainard Lake, 210-220 cm, stone scrape

no diatoms of stomatocysts seen

Appendix 3.1 (cont'd)

Whisler Lake, core #3, 273-284 cm, stone scrape

Planktonic forms	
Aulocoseira subarctica (O.Müller)Haworth	5
Undifferentiated forms	
Cymbella sp. (fragment)	2
Chrysophyte stomatocysts	3

Whisler Lake, core #1, 126 cm, drape layer

Planktonic forms	
Aulocoseira ambigua (Grun. in VanHeurck) Krammer	4
Aulocoseira subarctica (O.Müller)Haworth	4
Aulocoseira sp.	3
Cyclotella bodanica var. *affinis* Grun.	1
Stephanodiscus alpinus Hust.	2
Stephanodiscus cf. *minutulus* (Kütz)Round	5
centric sp. (unkown)	1
Littoral forms	
Cocconeis sp.	1
Tabellaria sp.	1
Undifferentiated forms	
Achnanthes minutissima Kütz	6
Amphora pediculus (Kütz)Grun.	2
Fragilaria brevistriata Grun.	3
Fragilaria construens v. *venter* (Ehr)Grun	14
Fragilaria pinnata Ehr.	17
Fragilaria pinnata var. *intercedens* (Grun.)Hust.	2
Fragilaria spp. (unknown)	9
Navicula cryptocephala Kütz.	3
Nitzschia sp.	3
Chrysophyte stomatocysts	5

TERTIARY OUTLIERS
AND REGIONAL LANDSCAPE EVOLUTION
OF EASTERN HAZEN PLATEAU,
NORTHERN ELLESMERE ISLAND, NUNAVUT, CANADA

4.1 INTRODUCTION

Christie (1974, 1976) and Miall (1979, 1991) have reported Tertiary outliers within Lake Hazen Basin and Hazen Plateau, northern Ellesmere Island (Figs. 4.1 and 4.2). Observations by this author add to the understanding and extent of these deposits, as well as the geological evolution of the region. Emphasis is placed upon the formation of the fiords and major valleys, and their relationship to the regional glacial history.

4.1.1 Regional Geology

Hazen Plateau consists of rocks belonging to the Danish River Formation, deposited in a deep marine basin during the early to mid-Paleozoic (Silurian-Devonian; Trettin 1994). Deformation and uplift occurred during the mid-Paleozoic Ellesmerian Orogeny after which Hazen Plateau underwent extensive peneplanation (Thorsteinsson & Tozer 1970). Trettin (1971) has suggested that subsequently, Hazen Plateau may have been covered by late Paleozoic (Permian) and Mesozoic sediments, because outcrops of such sediments occur along the Lake Hazen Fault Zone (Fig. 4.1). If so, their absence from regions south and east of Lake Hazen, reflects extensive erosion, prior to renewed Tertiary sedimentation (Trettin 1971).

Erosional remnants of Tertiary rocks unconformably overlie the Danish River Fm. southeast of Lake Hazen (Christie 1964, 1974, 1976; Miall 1979, 1991). Christie (1974) identifies these as numerous, small isolated knolls, composed of white sand beds with interbedded coal or coaly shale, and occasional petrified wood, but provides only a generalized map of their distribution. Miall (1979) more clearly delineates the region, between Salor Creek and Black Rock Vale (Fig. 4.2), where weakly-consolidated sandstone-mudstone members of Eocene age occur. Their deposition is attributed to small, high sinuosity, mixed load streams. No section log, or description of these outliers is provided, other than the noted presence of an abundant and well-preserved microflora in two samples bordering Salor Creek (Fig 4.2-site a, e; Miall 1979).

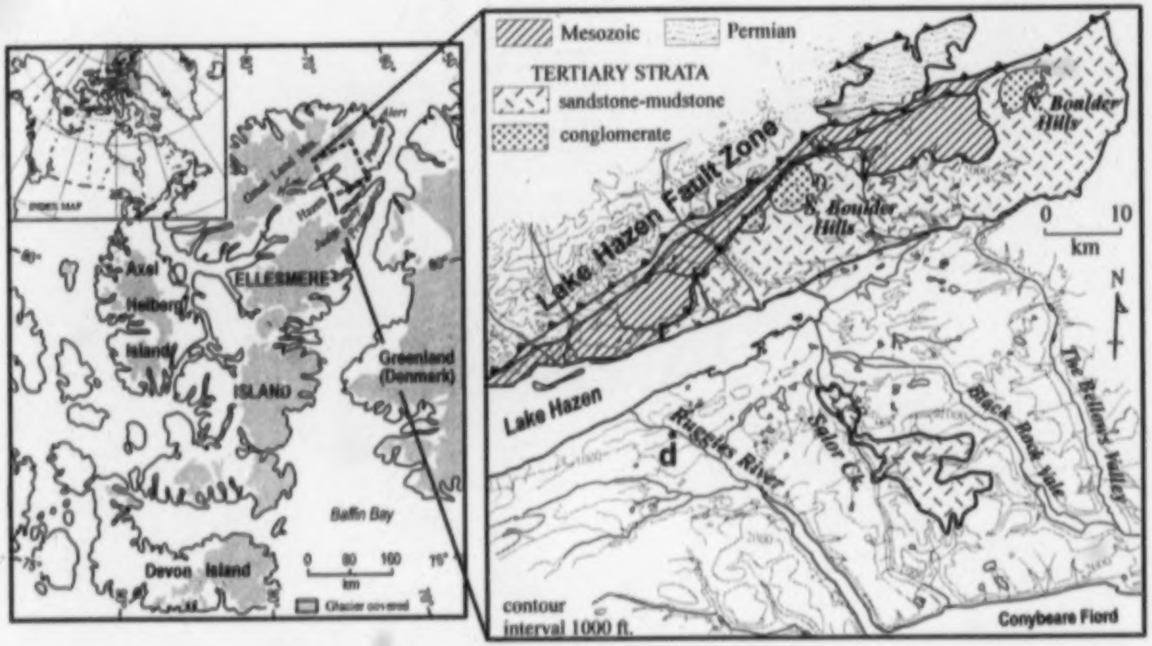
INDEX MAP
Axel
Heiberg
Island
ELLESMERE
ISLAND
Greenland
(Denmark)
Baffin Bay
Devon Island
0 80 160
km
Glacier covered
Mesozoic
Permian
TERTIARY STRATA
sandstone-mudstone
conglomerate
Lake Hazen Fault Zone
N. Boulder Hills
S. Boulder Hills
0 10
km
N
Lake Hazen
Ruggles River
Salor Ck.
Black Rock Vale
The Bellows Valley
d
contour
interval 1000 ft.
Conybeare Fiord

Figure 4.1 Maps showing the location and topography of the study area on northern Ellesmere island. Right hand map illustrates the distribution of Tertiary and Mesozoic strata identified by Miall (1979; redrawn with permission of the Geological Survey of Canada). Site **d** identifies the location of pre-Tertiary stratum along the Ruggles River (discussed in text).

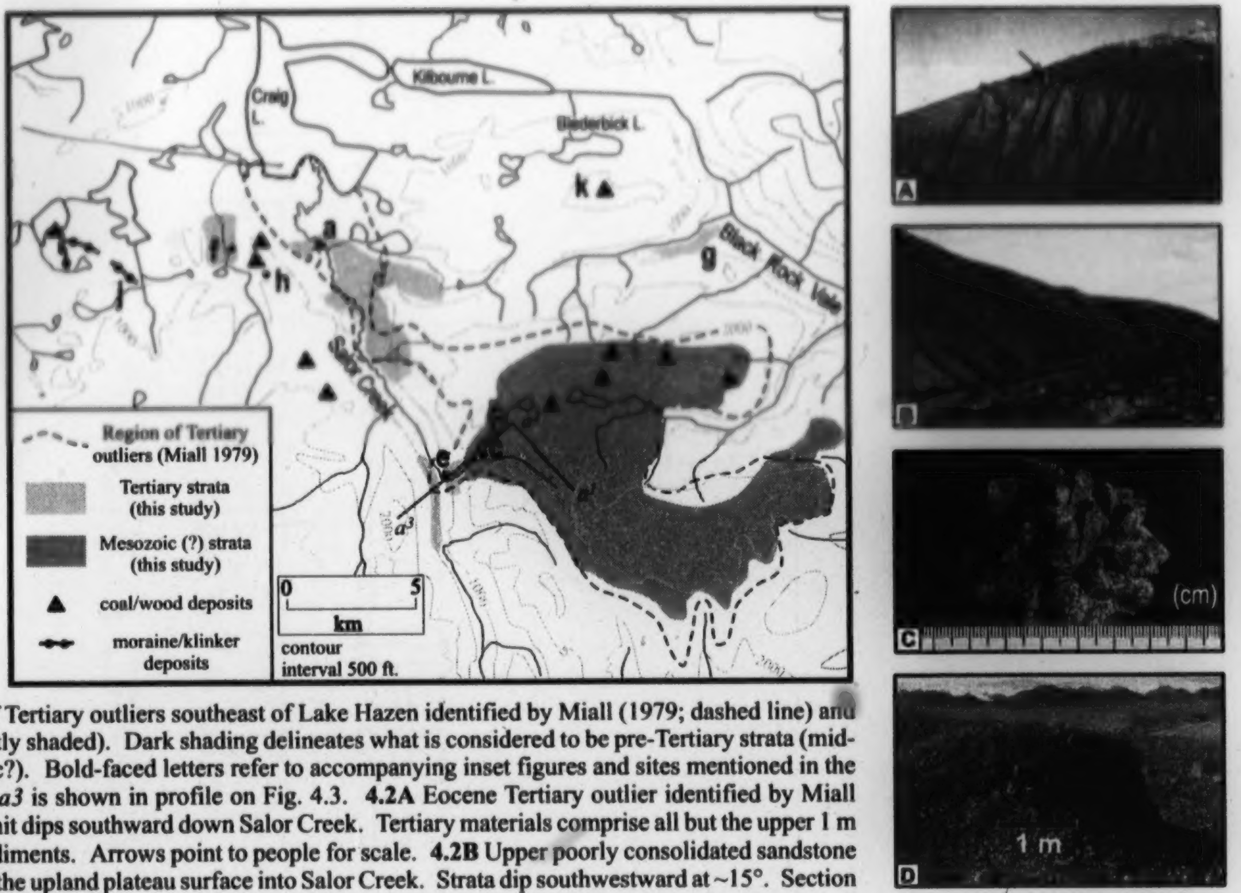
Kilbourne L.
Craig L.
Biederbick L.
k
a
g
h
Black Rock Vale
Salor Creek
e
a^3
a^2
Region of Tertiary outliers (Miall 1979)
Tertiary strata (this study)
Mesozoic (?) strata (this study)
coal/wood deposits
moraine/klinker deposits
0
5
km
contour interval 500 ft.
2000
1000
A
B
C
(cm)
D
1 m

Figure 4.2 Region of Tertiary outliers southeast of Lake Hazen identified by Miall (1979; dashed line) and that of this study (lightly shaded). Dark shading delineates what is considered to be pre-Tertiary strata (mid-Paleozoic to Mesozoic?). Bold-faced letters refer to accompanying inset figures and sites mentioned in the text. Transect *a1-a2-a3* is shown in profile on Fig. 4.3. **4.2A** Eocene Tertiary outlier identified by Miall (1979, his site 57). Unit dips southward down Salor Creek. Tertiary materials comprise all but the upper 1 m of pale Quaternary sediments. Arrows point to people for scale. **4.2B** Upper poorly consolidated sandstone unit, prograding from the upland plateau surface into Salor Creek. Strata dip southwestward at ~15°. Section is ~40 m high. **4.2C** Stellate nodule, a carbonate concretion indicative of cold marine waters, taken from creek bank exposure. **4.2D** Presumed remnant outlier of mid-Paleocene to early Eocene fluvial gravel deposit. Clasts are imbricate southward down the Ruggles River valley, and encased within a poor to well-lithified sandstone matrix (Fig. 4.1, site d).

4.2 FIELD EVIDENCE

4.2.1 Hazen Plateau

A discontinuous, thin, and recessive stratum (predominantly mudrock) unconformably overlies the Danish River Fm. on the upland Hazen Plateau (>600 m asl) between Salor Creek and Black Rock Vale (Figs. 4.2 and 4.3). This unit occurs within the region of Tertiary outliers delineated by Christie (1976) and Miall (1979, 1991), however, in the absence of a detailed description, its Tertiary designation appears to depend solely on its superposition atop the Danish River Fm. Several factors and observations by this author question its Tertiary designation. First, its upland setting, well above other horizontally-bedded Tertiary strata along the bottom of Salor Creek (Fig. 4.2, sites a, e), appears inconsistent with deposition by regional rivers, or within an intermontane lake basin (cf. Miall 1979). Second, the stratum overlying the Danish River Fm. does not extend from the valley floor up the north facing slopes of the plateau as depicted by Miall (1979; Fig. 4.2). Instead, the horizontally-bedded sandstone and mudrock overlying the Danish River Fm. along Salor Creek (Fig. 4.2A), is separated from the two or more strata overlying the Danish River Fm. atop the plateau (Figs. 4.2 and 4.3). Of these upper strata, the most extensive unit is an easily weathered, horizontally-bedded mudrock, which mantles much of the northern plateau surface (~600-880 m asl) between Black Rock Vale and Salor Creek (Figs. 4.2 and 4.3). Towards the headwaters of Salor Creek (Fig. 4.2, site b) a second unit overlies the Danish River Fm. (654-694 m asl). It is a poorly consolidated sandstone (containing well-consolidated interbeds), approximately 40 m thick, with horizontal to low angle bedding (<15°) which progrades to the upland surface, recording drainage from the northeast into the Salor Creek valley (Figs. 4.2B and 4.3). Downslope, in the large side-valley incised into the Danish River Fm., there are other discontinuous and onlapping strata (560-400 m asl; Fig. 4.3). They consist of poorly consolidated sediments with well-consolidated interbeds, and display near-horizontal to sharply deformed bedding (likely indicating down-valley slumping). Within the lower bedded sediments (~445 m asl; Fig. 4.2, site c), abundant stellate nodules (a calcite pseudomorph, also known as Polar Euhedrons; Fig 4.2C), and other large carbonate concretions were found. Stellate nodules have only been reported from Arctic Canada in Sverdrup Basin sediments of late Upper Jurassic to early Lower Cretaceous

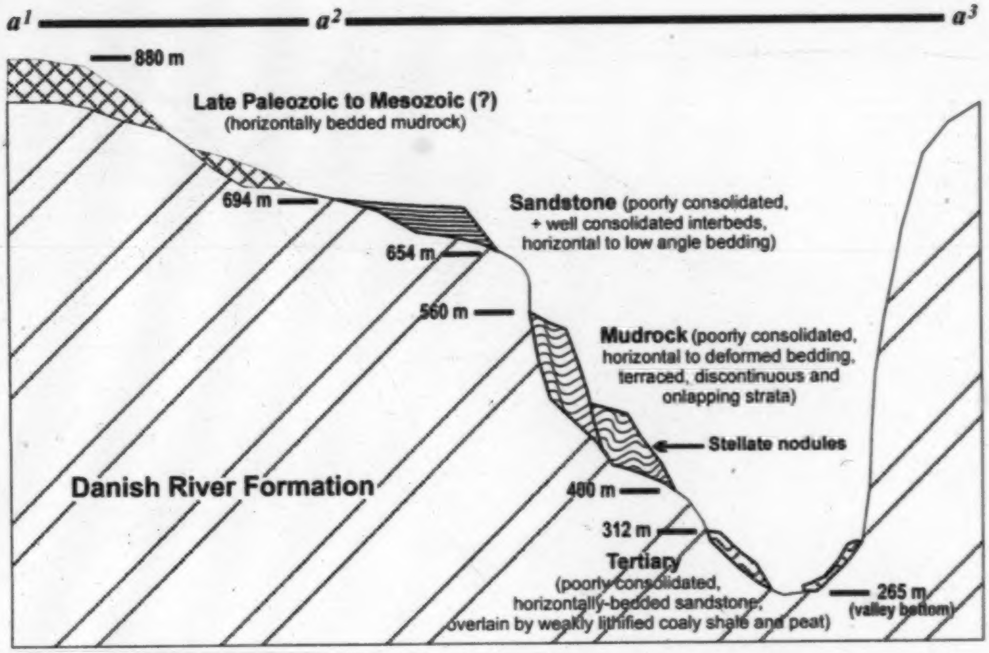
a1
a2
a3
880 m
Late Paleozoic to Mesozoic (?)
(horizontally bedded mudrock)
694 m
Sandstone (poorly consolidated,
+ well consolidated interbeds,
horizontal to low angle bedding)
654 m
560 m
Mudrock (poorly consolidated,
horizontal to deformed bedding,
terraced, discontinuous and
onlapping strata)
Stellate nodules
Danish River Formation
400 m
312 m
Tertiary
(poorly consolidated,
horizontally-bedded sandstone,
overlain by weakly lithified coaly shale and peat)
265 m
(valley bottom)

Figure 4.3 Cross-section sketch along the transect a^1 - a^2 - a^3 (Fig. 4.2), illustrating the different strata and their relative position along Salor Creek and on top Hazen Plateau. Tertiary stratum appears to be confined to the lower valley (<312 m asl), existing on both sides of the valley wall, and extending to within ~2 m of the valley bottom. The age of this unit is Eocene (Miall 1979), indicating that subsequent fluvial incision or glacial overdeepening of the Salor Creek valley has not occurred, other than the removal of much of the presumed Tertiary fill.

(Mesozoic) age, and are considered to indicate cold, marine environments (Kemper & Schmitz 1975). These conditions are incompatible with the paleoenvironmental reconstructions and non-marine origin of the regional Tertiary strata discussed by Miall (1979, 1991). They do, however, closely resemble the Mesozoic history of the Arctic Islands discussed by Embry (1991). Glacial redeposition of the stellate nodules is unlikely, as the unit contains well-lithified interbeds and there is an absence of glacial erratics (coal and chert pebble conglomerate). Wood, though locally abundant on the surface, is also absent within the beds, and is likely derived from the overlying till veneer.

The structure and dip of the beds of the upper unit between 654 and 694 m asl (Fig. 4.3) suggest that the Salor Creek valley must have been considerably shallower than it is presently. Hence, subsequent, horizontally bedded strata found below, suggest a progressive uplift of the Hazen Plateau surface, and incision (fluvial?) of the Salor Creek valley. The sediment source for these lower deposits is uncertain, but their reduced size and areal extent, argues against regional rivers draining from atop the upland plateau. Thus, the continued incision of the Salor Creek valley, may coincide with disruption of regional river systems, and the initiation of other regional valleys (i.e. Black Rock Vale, The Bellows Valley) which presumably did not exist at the time rivers crossing Hazen Plateau deposited this upper unit.

4.2.2 Tertiary Outliers

Continued uplift of the Salor Creek valley from a marine to a subaerial environment must have occurred, as Miall (1982, 1991) identifies the Grant Land Mountains and Hazen Plateau as a sediment source for early to mid-Paleocene fluvial deposits on Judge Daly Promontory (Fig. 4.1). This deposition would require that Archer Fiord had not yet formed. Continued uplift and tilting of Hazen Plateau occurred during the late-Paleocene to early Eocene, as Miall (1979) records radial paleocurrents depositing the Tertiary strata northeast of Lake Hazen (Fig. 4.1), in what is interpreted to be an intermontane basin. Horizontally-bedded Tertiary (Eocene) strata in the bottom of the Salor Creek valley (Fig 4.2, sites a, e), are interpreted to indicate a shallow lake environment (Miall 1979). This confirms the observations of this author of poorly lithified sandstone interbedded with well-lithified mudstone (Fig. 4.2A, dip is southwards up Salor Creek) and other horizontally-bedded strata along Salor Creek. These include thin (10-20 cm) coaly shale beds overlain by poorly lithified, organic-rich (peat?) deposits in the side-valley (occurs just above Miall's 1979, site 58 = Fig 4.2, site e, ~312 m asl), suggesting a progressive shallowing of the lake.

Poor to well-lithified, horizontally-bedded strata containing abundant, well-lithified wood (ranging from small pieces up to 30x40 cm logs) was also observed at site f (Fig. 4.2), outside the previously mapped region of Tertiary strata. Based on the presence of wood, this unit is tentatively assigned a Tertiary age, but remains to be studied further. A small, horizontal bedrock terrace overlying the Danish River Fm. at the head of Black Rock Vale (Fig. 4.2, site g) is also tentatively grouped with the Tertiary strata.

There are several sites where there are localised deposits of coaly shale, coal and wood (of presumed Tertiary age; Fig. 4.2 - represented by triangles) which closely match the description and delineation of Tertiary outliers by Christie (1974, 1976). However, field observations from several sites suggest that many or all of these are glacially redeposited rafts, or kames, rather than being *in situ* deposits. Reasons for this interpretation include: the observation of till overlying and/or underlying several deposits; large beds of coal that are contorted and dipping vertically (Fig. 4.2, site h); extensive klinker deposits along a deglacial ice margin (Fig 4.2, site j); and the presence of well-lithified logs found on the surface around Craig and Biederbick lakes. Glacial rafting also provides one explanation for the extensive piles of well- to poorly lithified wood and coal along the north facing rim of the plateau between the Ruggles River and Black Rock Vale (Fig. 4.2). In particular, there is no evidence of a dispersal train leading southwards from these deposits, despite the fact that they are covered or associated with a thin till veneer. Site k (Fig 4.2), south of Biederbick Lake, also contains abundant fossilized wood (logs) lying on the surface, and a 50+ cm thick and 10 m long horizontal bed of coal overlain by till (lower contact is obscured).

A 3 m high, by 10+ m long, outcrop of well-lithified, horizontally bedded, sandstone, containing abundant current bedded structures (paleocurrents southwards down-valley), was also observed along the shore of the Ruggles River, directly overlying the Danish River Fm. (Fig. 4.1, site d; ~126 m asl). The age of this bed is unknown, but its marine nature (carbonate matrix) is incompatible with the regional Tertiary strata (Miall 1979, 1991), and therefore, is tentatively assigned a pre-Tertiary age. Thus, it similarly supports a pre-Tertiary age for the formation of the Ruggles River valley (modern river bottom is only 1 m below the base of the observed unit). Above this (156 m asl) are several small (10x6 m) outcrops of poorly consolidated, imbricate gravels and cobbles, supported by a well to poorly-lithified sandstone matrix (Fig. 4.2D). The depositional history of these latter units is uncertain, but may relate to the period of fluvial deposition from Hazen Plateau and Grant Land Mountains onto Judge Daly Promontory during the early to mid-Paleocene (Miall 1979, 1982).

4.3. DISCUSSION

4.3.1 Regional Valley and Fiord Formation

The observations reported here are important for two main reasons. The first concerns the record of valley incision, and the implications this has for the proposed Tertiary fluvial vs. tectonic origins of inter-island marine channels in the Arctic Archipelago (Fortier & Morley 1956; Trettin 1991). One hypothesis suggests that the inter-island channels were formed by large river systems, established during late stages of the Eurekan Orogeny (middle to late-Tertiary; Trettin 1991). River channels were proposed to have experienced subsequent glacial overdeepening, forming classical "U" shaped valleys (Pelletier 1966), and are regarded as both examples of selective linear erosion (Sugden 1977), and evidence that glacial erosion is responsible for the major elements of the Canadian Arctic landscape (Denton & Hughes 1981). An alternative hypothesis, discussed in England (1987), ascribes the interisland channels to widespread block faulting, of late-Tertiary origin (Kerr 1980; England 1987). Evidence supporting block faulting is probably best illustrated by the work of Bornhold *et al.* (1976) and Dyke (1983), in the southern Arctic Islands, which illustrates remnant upper-Paleozoic to Tertiary river channels on both the resulting islands (horsts) and on the floors of the intervening channels (grabens). On Hazen Plateau, the fact that the Salor Creek and Ruggles River valleys contain pre-Tertiary strata, indicates that the present-day valley forms, which crosscut obliquely the underlying Danish River Fm., predominantly record antecedent drainages, dating as early as the Miocene. Furthermore, the horizontally-bedded strata at the head of Salor Creek (Fig. 4.2. site a) are exposed right down to the bottom of the valley (marked by the contact with the underlying Danish River Fm.). Further upvalley, the valley bottom (265 m asl) is similarly floored by the Danish River Fm., and the lower Tertiary fill lies as little as 2 m above the valley floor. Hence, fluvial incision and glacial overdeepening of the Salor Creek valley during the post Eocene, have been minimal, removing ~40 m of poorly consolidated Tertiary valley-fill. At the south end of the Salor Creek valley, spectacular, linear cliffs (500 - 1000 m high) line the coastline, with water depths reaching between 200 and 600 m in Conybeare Fiord and adjacent Archer Fiord and Lady Franklin Bay. That these fiords were formed by late Tertiary fluvial incision, is contradicted by the nominal erosion in their adjacent tributaries which involved both fluvial and glacial erosion. Hence, the fiords must have a predominantly tectonic origin.

4.4 CONCLUSIONS

Tertiary outliers identified in the southeastern Lake Hazen Basin and Hazen Plateau by Christie (1964, 1976) and Miall (1979, 1991) do not entirely conform to recent field observations. Rather, the horizontally bedded mudrock atop Hazen Plateau, between Salor Creek and Black Rock Vale, likely represents late Paleozoic to Mesozoic marine strata deposited across a mid to late Paleozoic peneplain. A proto-Salor Creek valley must have formed during the later stages of mid to late-Paleozoic peneplanation, and existed as a marine embayment. The origin of the poorly consolidated sandstone unit overlying the mudrock and Danish River Fm. is uncertain (i.e., marine, deltaic, terrestrial), but may relate to continued uplift and fluvial dissection of Hazen Plateau. Poorly consolidated, horizontally-bedded sediment with well-lithified interbeds below these upper units, contains abundant stellate nodules. These rocks are associated with cold, marine waters (Kemper & Schmitz 1975) and are therefore incompatible with the regional Tertiary strata interpreted to have been deposited within an intermontane basin (Miall 1979, 1991). Other marine strata found along the Ruggles River, overtopped by what are interpreted to be outliers of Tertiary fluvial gravels, also support a pre-Tertiary age for the formation for this major valley. The extent of Tertiary outliers southeast of Lake Hazen is thus considerably less than that mapped by Christie (1976) and Miall (1979), and is generally confined to small discontinuous sections in the bottom of valleys (Fig. 4.2).

The presence of undisturbed, poorly consolidated, early Tertiary (Eocene) deposits only a couple of metres above the floor of Salor Creek argues that post-Eocene fluvial and glacial erosion at this site have been minimal. A similar argument is made for the Ruggles River valley, where a well lithified, current-bedded sandstone of presumed pre-Tertiary age, lies only one metre above the bed of the Ruggles River. Fluvial gravel remnants, which would have been deposited after this, and are likely correlative with the mid-Paleocene to early Eocene deposition from the Grant Land Mountains onto Judge Daly Promontory (Miall, 1982), are found 30 m higher along the west shore of the river. Erosion then (fluvial and glacial) has been expended removing the Tertiary fluvial sediments, and hence is not responsible for creating the larger valley form. It seems illogical then to suggest that excessive fiord depths of 200 - 600 m and linear cliffs (500-1000 m high) along Chandler, Conybeare and Archer fiords are the product of fluvial incision and glacial overdeepening postdating the Eocene. Instead, the argument that they are the product of late-Tertiary tectonics (block faulting) is more consistent with the sedimentary and geomorphic record.

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THE LATE QUATERNARY GLACIAL HISTORY OF LAKE HAZEN BASIN AND EASTERN HAZEN PLATEAU, NORTHERN ELLESMERE ISLAND, NUNAVUT, CANADA

5.1 INTRODUCTION

This study examines the late Quaternary glacial history of Lake Hazen Basin and eastern Hazen Plateau, northern Ellesmere Island (Fig. 5.1). It is distinguished from previous studies in that much of the investigated region occupies the broad interior, rather than the more extensively examined coastlines of Ellesmere Island. The location and topography of this region provide valuable insights into past glaciations and paleoenvironmental change, and records the interaction between glaciers emanating from the Grant Land Mountains, Hazen Plateau, Agassiz Ice Cap and Greenland (Fig. 5.1).

5.1.1 Previous Investigations

Based on the regional physiography, Taylor (1956) proposed that the Greenland Ice Sheet once crossed Hazen Plateau. Later fieldwork indicated that Greenland erratics occurred no more than 20 km inland of the eastern Hazen Plateau, demonstrating that a local (Grant Land Mountains) ice source infilled the basin (Smith 1961; Christie 1967). Subsequent investigations (England & Bradley 1978; Retelle 1986) confirmed Christie's observations on the distribution of Greenland erratics (observed up to 670 m above sea level (asl); Fig. 5.1), and documented the prior existence of plateau ice caps and stratigraphic evidence for two or more glaciations. England (1976a, 1978, 1983, 1996) documented the glacial and raised-marine record throughout Archer Fiord, Lady Franklin Bay, and Judge Daly Promontory. From this, he suggested that the LGM was characterised by only moderate advances of local ice caps (5-60 km) beyond present margins, supplemented by the growth of local plateau ice caps. Glaciers were considered constrained by aridity and their inability to overcome fiord depths (cf. Lemmen *et al.* 1994). A prominent, discontinuous belt of low elevation lateral moraines (100-300 m asl), situated at many fiord heads, was considered to mark the limit of the last glaciation (England 1978). Dubbed the "Hazen Moraines" they represented a chronostratigraphic margin, similar to the "drift belt" proposed by Hodgson (1985) on the west coast of Ellesmere Island, and recorded the contact of glaciers with a "full-glacial sea" ca. 8-8.5 ka BP (England 1983). The Hazen Moraines also included conspicuous moraines ringing the east end of Craig Lake, southeast of Lake Hazen (although these remained undated; Fig. 5.1). A similar discontinuous belt of moraines, marking a

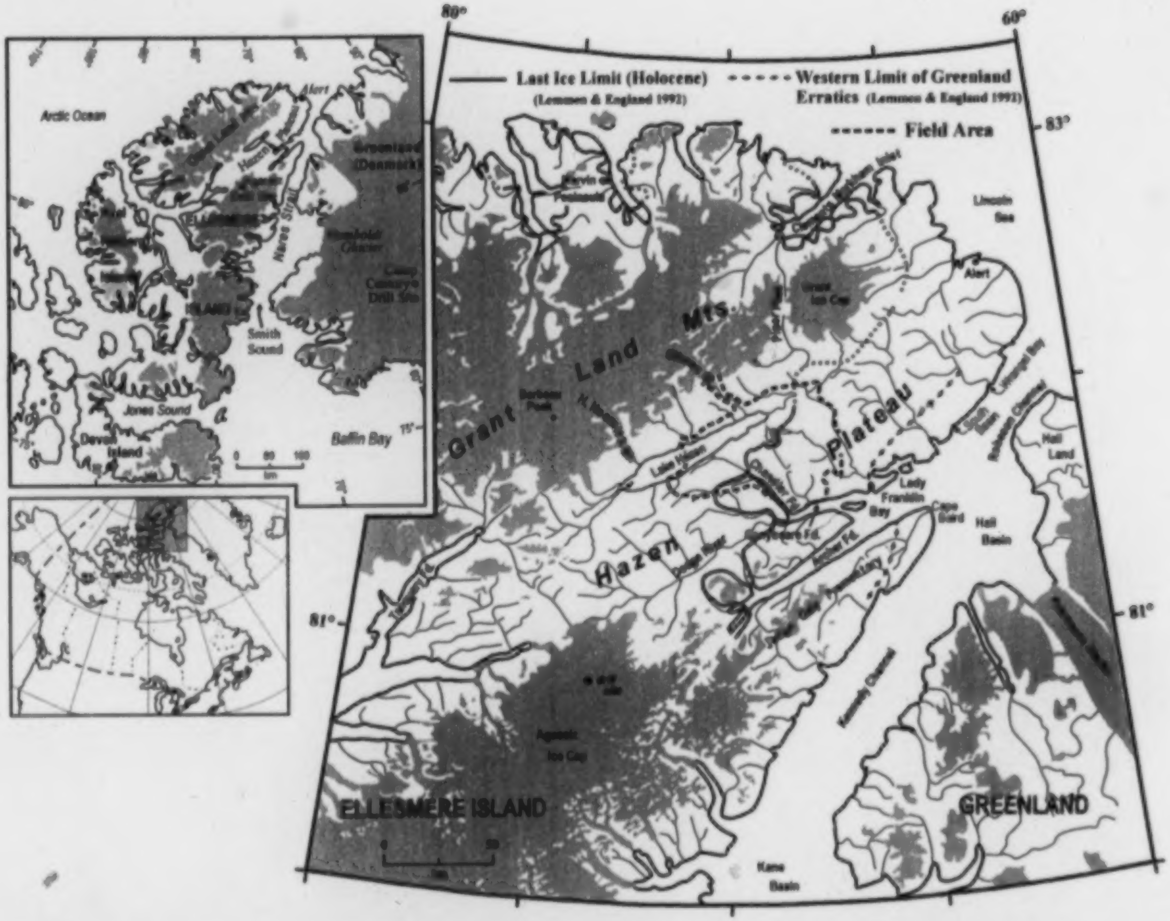

80°
60°
83°
81°
81°
Last Ice Limit (Holocene)
(Lemmen & England 1992)
Western Limit of Greenland Erratics (Lemmen & England 1992)
Field Area
Arctic Ocean
Alert
Greenland (Denmark)
Nares Strait
Humboldt Glacier
Camp Century Drill Site
Smith Sound
Jones Sound
Devon Island
Baffin Bay
ELLESMERE ISLAND
Grant Land Mts.
Hazen Plateau
Lake Hazen
Lincoln Sea
Lady Franklin Bay
Cape Baird
Hall Basin
Hall Land
Kennedy Channel
Kane Basin
Agassiz Ice Cap
GREENLAND

Figure 5.1 Location maps of the northeastern Canadian Arctic Archipelago, showing detail of the principal field area on northeastern Ellesmere Island. Shading depicts the contemporary glacier cover.

limited advance of ice, was noted by Bednarski (1986) in Clements Markham Inlet (Fig. 5.1), and by other researchers along the northern Ellesmere Island coast (Lemmen 1989; Evans 1990). The proposed restricted ice margins at the LGM also appeared to support the hypothesis that large areas of Hazen Plateau constituted a biological refugium during the last glaciation (Leech 1966; Brassard 1971; Røen 1981).

Beyond the Hazen Moraines, the geomorphic and chronological evidence of previously more extensive glaciations suggested that they dated >35 ka BP (England & Bradley 1978; England *et al.* 1981; Lemmen & England 1992). Retelle (1986), working along the eastern Hazen Plateau adjacent Robeson Channel, identified what was believed to be a marine shoreline at 286 m in Wrangel Bay, below which glaciomarine silts contained shells which dated >30 000 BP (considered a minimum age). Amino acid analyses of these same shells (Total and Free allo-isoleucine and isoleucine ratios) were interpreted to be disparate from Holocene deposits, and were ascribed to the Robeson aminozone, dating ≥70 ka, considered a limiting date on the penultimate glaciation (Retelle 1986).

On northwestern Greenland, England (1985, 1987a) proposed a similarly limited advance of ice at the LGM where ice margins calved in a full-glacial sea occupying Nares Strait (the collective name for the channels separating Ellesmere Island and Greenland - Fig. 5.1). Contrasting with this, was the view that during the LGM Greenland ice crossed Nares Strait and impinged on northeastern Ellesmere Island (Weidick 1975; Bennike *et al.* 1987; Kelly & Bennike 1992; Funder & Hansen 1996). Further south, Greenland and Ellesmere ice were considered to have coalesced in Kane Basin, flowing southward through Smith Sound into Baffin Bay (Blake 1992, Blake *et al.* 1992; Blake *et al.* 1996). Previous, more extensive glaciations were considered responsible for high elevation erratics and glacial geomorphology, which might be correlative with the deposition of the highest Greenland erratics on Ellesmere, and the Robeson aminozone (Kelly & Bennike 1992).

5.1.1.1 *Recent Revision of the Quaternary History*

England (1998, in press) has fundamentally revised his earlier work, proposing that Greenland and Ellesmere ice were coalescent throughout Nares Strait at the LGM. Furthermore, England (1998, in press) proposes that the highest elevation Greenland erratics on Judge Daly Promontory, and by inference the outer Hazen Plateau, were deposited during the last glaciation. This reinterpretation is based on geomorphic evidence for trunk ice in Nares Strait and radiocarbon dates on ice-contact deltas documenting its retreat southward from Alert and northwards from Smith Sound. All previous evidence used to argue against the occupation of Nares Strait during the LGM (i.e., weathering zones, pre-Holocene shorelines, and the presence of a full-glacial sea) has been rejected (England, in press). The interpretation that the Robeson aminozone records the last retreat of Greenland ice off Ellesmere Island (≥70 ka BP; Retelle 1986) was also shown to be incorrect because younger ratios can be found on shells in till deposited by the Nares Strait trunk glacier (England, in press).

5.2 STUDY AREA

5.2.1 Physiography and Climate

The area is characterised by three physiographic regions: Grant Land Mountains, Lake Hazen Basin, and Hazen Plateau (Fig. 5.1). The Grant Land Mountains rise abruptly from the north shore of Lake Hazen (157 m asl), reaching 2665 m asl (Barbeau Peak). They support a large ice cap (up to 900 m thick), from which trunk glaciers descend to 560 - 240 m asl above Lake Hazen, and to sea level along the island's north coast (Hattersley-Smith *et al.* 1969; Narod & Clarke 1983). The mountains trend SW-NE along the Lake Hazen Fault Zone, and are formed of Paleozoic limestones, quartzose and feldspathic sandstones, slate and chert pebble conglomerate (Christie 1964; Trettin 1994). This chert pebble conglomerate outcrops in nunataks along the central and southern Grant Land Mountains (Christie 1964) and is a conspicuous glacial erratic.

Lake Hazen Basin comprises the lowland areas (<300 m asl, ~3500 km^2) encircling Lake Hazen. East of Lake Hazen Basin, Hazen Plateau (9250 km^2) rises to 1300 m asl, contacting the sea in spectacular linear cliffs along Archer Fiord, Lady Franklin Bay and Robeson Channel. The principal surface rock of Hazen Plateau is the Danish River Formation (Trettin 1994; formerly the Imina Fm., Trettin 1971) which consists locally of >620 m of calcareous and dolomitic sandstone and mudrock, tightly folded along a pronounced SW-NE axial plane. Overlying the Danish River Fm. is a wedge, up to 600 m thick, of Upper Paleozoic to Tertiary beds, that extends northeastward from Lake Hazen for approximately 80 km. Tertiary erosional remnants are also found in eastern Lake Hazen Basin (Christie 1964, 1974; Miall 1979, 1991; this thesis Chapter 4). Tertiary units are comprised of weakly lithified sandstone, sandy shale, ironstone, coal, amber, and carbonized and silicified wood.

Large, incised valleys and fiords crosscut the Danish River Fm., at angles oblique and perpendicular to strike. The origin of these is ascribed to faulting and antecedent drainages possibly dating back to the mid- to late-Paleozoic (this thesis Chapter 4). The presence of poorly consolidated Eocene units only a couple of metres above the base of Salor Creek and Ruggles River, requires that post-Eocene incision (fluvial and glacial) has been minimal. Linear cliffs (600-1000 m high) and deep fiords (200-600 m) are likely the product of regional late-Tertiary block faulting (England 1987b), and not as suggested by some, post-Eocene Tertiary river and glacial incision (Fortier & Morley 1956; Miall 1979; Trettin 1991).

Lake Hazen Basin and surrounding region are polar deserts (Bovis & Barry 1974), with a mean annual temperature of approximately -20°C, and a mean annual precipitation of <150 mm water equivalent (Thompson 1994). Owing to the continentality and topography of the region, it becomes a "thermal oasis" in the brief summer, and temperatures can exceed 18°C. Cold-air drainage and severe temperature inversions similarly account for mean winter monthly temperatures (~ -40°C) that are 6 to 10°C colder than Alert (Thompson 1994).

5.3 RESEARCH METHODS

Reconstruction of the late Quaternary glacial history involved mapping of glacial erosional and depositional features, surveying of raised marine and lacustrine sediments, and coring of modern lake basins. Mapping was based on air photo interpretation of much of Hazen Plateau, and intensive field investigations south and east of Lake Hazen (Fig. 5.1). Former ice margins, delineated by moraines, kames, meltwater channels, glaciomarine deltas and proglacial lake deposits, were mapped and surveyed by altimetry. Altimeter readings were corrected for changes in pressure and temperature, and each base camp was calibrated to a datum of Lake Hazen (157 m asl). However, long foot traverses over relief >500 m, and an absence of benchmarks, resulted in errors of ±5 m. Whenever possible, key sites were repeatedly surveyed, reducing altitudinal errors to ±2 m. Former ice profiles were measured by tracing lateral meltwater channels on the ground using altimetry, and from air photographs by tracing channels (2-4 km long) which were then transposed to 1:50 000 topographic maps (10 m contour interval). Sixteen extant lake basins, eight of which lay beyond the formerly proposed limit of the last glaciation (Fig. 5.1; England 1983), were cored in an attempt to retrieve preglacial sediments. Chronological control was provided by Accelerator Mass Spectrometry (AMS) radiocarbon dating of organic materials in lacustrine and marine deposits, and by ^{36}Cl surface exposure dating of erratics and bedrock (Tables 5.1 and 5.2).

5.4 FIELD OBSERVATIONS

5.4.1 Glacial Geomorphology

The surficial geology is characterised by bedrock overlain in places by felsenmeer, or till veneer. Moraines are rare, but do record the retreat and southward flow of outlet glaciers from a Grant Land Mountain trunk glacier occupying Lake Hazen Basin. Other glacial depositional landforms are not widespread, with the exception of those associated with proglacial lakes (impounded between retreating glaciers and the rising plateau). These include extensive blankets of glaciolacustrine sediment and isolated deposits of rhythmites and current-bedded sand and silt up to 32 m thick.

Lateral meltwater channels and spillways from proglacial lakes are by far the most prominent erosional landforms, particularly west of the Ruggles River (Fig. 5.2a) and northeast of The Bellows Valley (Fig. 5.2b). In some areas, dense networks of channels, some >50 m deep, dissect the bedrock. *En echelon* patterns of lateral meltwater channels record the thinning and progressive topographic control over the retreat of a formerly pervasive ice cover, into regionally-confined lobes (Fig. 5.3).

Isolated and nested rows of kames up to 21 m high, are prominent glacio-depositional features bordering the Ruggles River (Fig. 5.4). More than 100 kames were observed in 150 km^2 west of the Ruggles River. Kames are common features along the termini of modern glaciers in the region where they form from deeply entrenched, perennial, supraglacial meltwater channels, that incise debris-rich thrust faults exposed along the near-vertical

Table 5.1 Radiocarbon date list. Dates are reported with one standard error, except GSC samples which are reported to two standard errors. Organic materials were pretreated with an acid-alkali leach, shells were leached in acid (R. Beukens, pers. comm. 1996). Shell samples have been corrected for a reservoir age of 410 years.

Location	Lab. dating #	Age (years BP)	Material	Enclosing sediment	Sample elevation (m asl)	Related relative sea level (m asl)	Lat. (N)	Long. (W)	Reference
1. Lake Hazen Basin	TO-3488	43 510±690	*Hippuris vulgaris* + *Potomogetan* seeds	lacustrine silt	160		81°57'	68°55'	This study
2. M[illegible]Creek	TO-4205	37 260±470	*Cassiope tetragona*	lacustrine silt	350		81°54'	69°47'	This study
3. Clements Markham Inlet	S-2140	23 850±850	Salix twig	dedrital, stratified sand	85		82°42'	68°07'	Bednarski 1986
4. Alert	GSC-1815	10 100±210	shells, *in situ*	silt	~100	120	82°27'	62°40	England 1976b
5. James Ross Bay	S-1984	9825±460	shells, whole valves	silt and sand	85	≥90	82°42'	64°45'	England 1983
6. Clements Markham Inlet	S-1981	9745±255	shells, *in situ*	silt	100	117	82°38'	67°34'	Bednarski 1986
7. Central Hall Land	GSC-3744	9580±140	shells, *in situ*	silt	110	≥110 - ≤140	81°42'	59°36'	England 1985
8. Wood River	S-1985	9270±1055	shells, *in situ*	bottomset silt	~45	≥92 - ≤125	82°30'	68°15'	England 1983
9. Clements Markham Inlet	S-2105	9270±265	shells, *in situ*	silt	87	111	82°43'	67°57'	Bednarski 1986
10. East Newman Bay	S-2307	9070±150	shells, whole valves	silt	92	>92	81°49'	58°40'	England 1985
11. Clements Markham Inlet	S-2211	8915±115	driftwood	surface	84	>84	82°38'	68°00'	Bednarski 1986
12. Lincoln Bay	SI-5550	8600±90	shells, *in situ*	silt	91	100	82°08'	62°03'	Retelle 1986
13. Archer Fiord	TO-136	8520±80	shells	beach	107	≥107 - ≤115	81°23'	66°53'	England in press
14. Piper Pass	S-2501	8415±135	caribou antler	silt	25		82°36'	68°31'	Stewart & England 1986
15. Cape Baird terrace	DIC-737	8380±105	shells, *in situ*	topset beds	~105	≥110 - ≤120	81°33'	64°30'	England 1985
16. Beaufort Lakes	S-1990	8255±215	shells, *in situ*	silt	90	116	81°53'	63°20'	England 1983
17. Ida Bay	GSC-1775	8130±200	shells, *in situ*	silt		104	81°32'	68°58'	England 1978
18. Chandler Fiord	GSC-3179	7860±270	shells	silt	65	83 (≥87)	81°41'	69°08'	England 1983

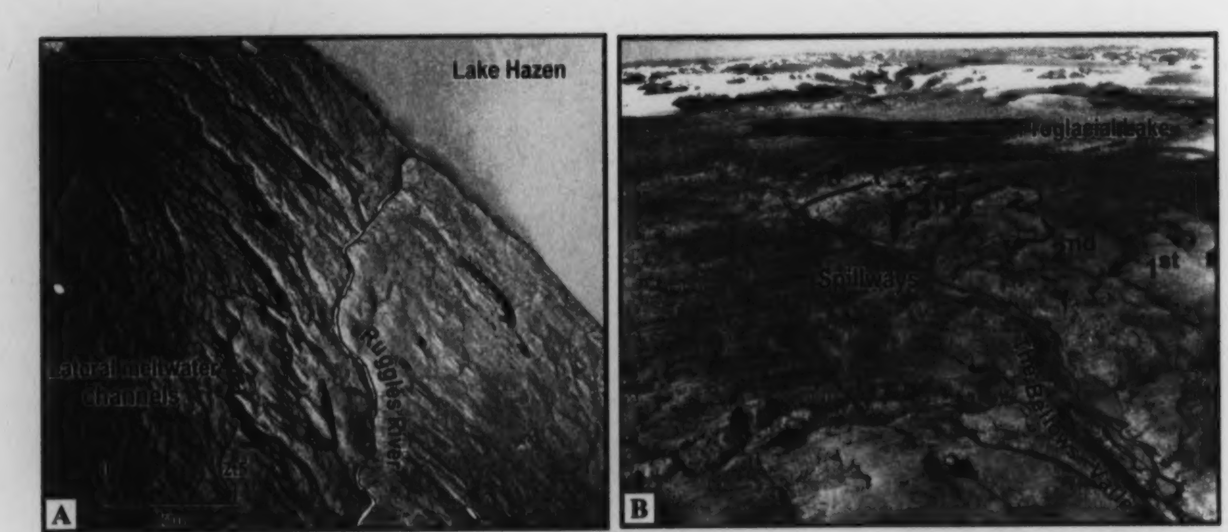
Table 5.1 (continued).

Location	Lab. dating #	Age (years BP)	Material	Enclosing sediment	Sample elevation (m asl)	Related relative sea level (m asl)	Lat. (N)	Long. (W)	Reference
19. Fort Conger	TO-435	7830±90	shells, *in situ*	bottomset clay	33	>62 - ≤114	81°45'	64°47'	England 1997
20. Clements Markham Inlet	GSC-2975	7830±80	driftwood	surface sample	72	>72	82°36'	68°25'	Stewart & England 1983
21. Sun Cape	S-2109	7755±125	shells	surface sample	102-105	>105 - ≤114	81°40'	65°20'	England 1983
22. Whisler Lake	TO-4473	7600±400	bryophytes	lake sediment	(80.5 cm core depth)		81°54'	68°24'	This study
23. Brainard Lake	TO-5274	7550±80	bryophytes	lake sediment	(141 - 144 core depth)		81°45'	68°11'	This study
24. Chandler Fiord	TO-4474	7550±90	shells, *in situ*	silt	58	≥87	81°41'	69°08'	This study
25. Muskox Bay	TO-436	7390±80	shells, *in situ*	bottomset silt	70	>70 - ≤110	81°46'	65°43'	England 1997
26. The Bellows Valley	S-2139	7385±375	shells	bottomset sand	32	≥37	81°41'	66°21'	England 1983
27. Black Rock Vale	S-2110	6995±130	shells	silt	50	≥70	81°47'	67°37'	England 1983
28. Sun Bay	SI-4762	6750±80	shells	silt	78	>78	81°38'	66°10'	England 1983
29. Black Rock Vale	TO-4475	6690±90	shells	topsets	53	55	81°47'	67°13'	This study
30. Piper Pass	TO-284	6410±50	bryophytes	peat	450		82°11'	68°28'	LaFarge-England *et al.* 1991
31. Appleby Lake	TO-5149	6290±80	bryophytes	lake sediment	(102 cm core depth)		81°51'	68°16'	This study
32. Chandler Fiord	S-1980	6255±110	shells, *in situ*	foreset beds	62	72	81°42	69°15'	England 1983
33. Chandler Fiord	St-4087	6100±260	shells, *in situ*	foreset beds	~60	72	81°43'	69°15'	England 1978
34. Craig Lake moraine	TO-4206	5970±70	bryophytes	peat	176		81°52'	68°33'	This study
35. Connell Lake	TO-5867	5290±110	bryophytes	sorted sediment	(168 - 170 cm core depth)		81°47'	69°15'	This study
36. Skeleton Creek	GSC-3451	4980±70	peat	peat	230		81°49'	71°18'	Blake 1986
37. Eugene Glacier	GSC-1813	4970±60	peat	peat	610		82°17'	66°18'	Lowdon & Blake 1979

Table 5.2 Cosmogenic ^{36}Cl date list.

Site	Material	Age (ka)[a]	Elevation (m asl)	Lat. (N)	Long. (W)
1. Chandler Moraine	striated boulders	18.5 ± 3.1	622	81°38'	68°43'
2. Hazen Plateau	bedrock	14.3 ± 4.1	310	81°42'	70°17'
3. Craig Lake Moraine	striated boulder	9.8 ± 1.5	235	81°52'	68°35'
4. Black Rock Vale Moraine	striated boulders	8.0 ± 0.5	515	81°53'	67°28'

[a] Age is reported in sidereal years. Samples were analysed, and dates provided by Dr. M. Zreda, University of Arizona (pers. comm. 1996)


Lake Hazen
Ruggles River
Lateral meltwater channels
A
Proglacial lake
2nd
1st
Spillways
The Bellows Valley
B

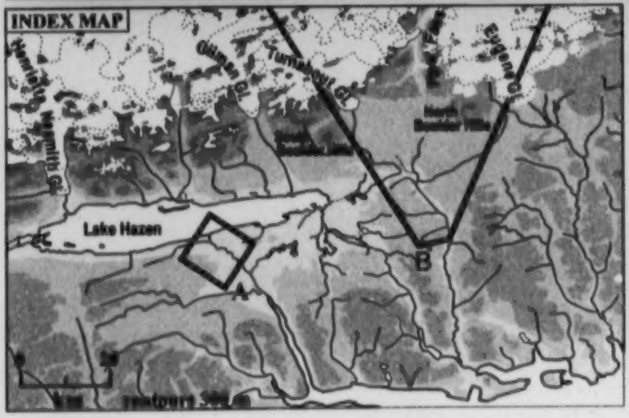

INDEX MAP
Henrietta Nesmith Gl.
Gilman Gl.
Turnabout Gl.
Eugene Gl.
North Boulder Hills
Lake Hazen
A
B
contours 300 m

Figure 5.2 Prominent glacial erosional features. **5.2A** Vertical air photograph (A16607-82 (© 1959)) illustrating the intense dissection of terrain west of the Ruggles River by lateral meltwater channels (barbs point up-slope). Ice retreated northwards towards Lake Hazen. **5.2B** Oblique air photograph (T490L-132 (© 1952)) shows spillways from a large proglacial lake, formed at the head of The Bellows Valley when the trunk glacier blocked the regional drainage of the Eugene Glacier and western Grant Ice Cap. Spillways record the sequential lowering of the lake, accordant with the northward retreat of ice. **Index Map** shows the location of the two figures. [Air photographs © Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada]

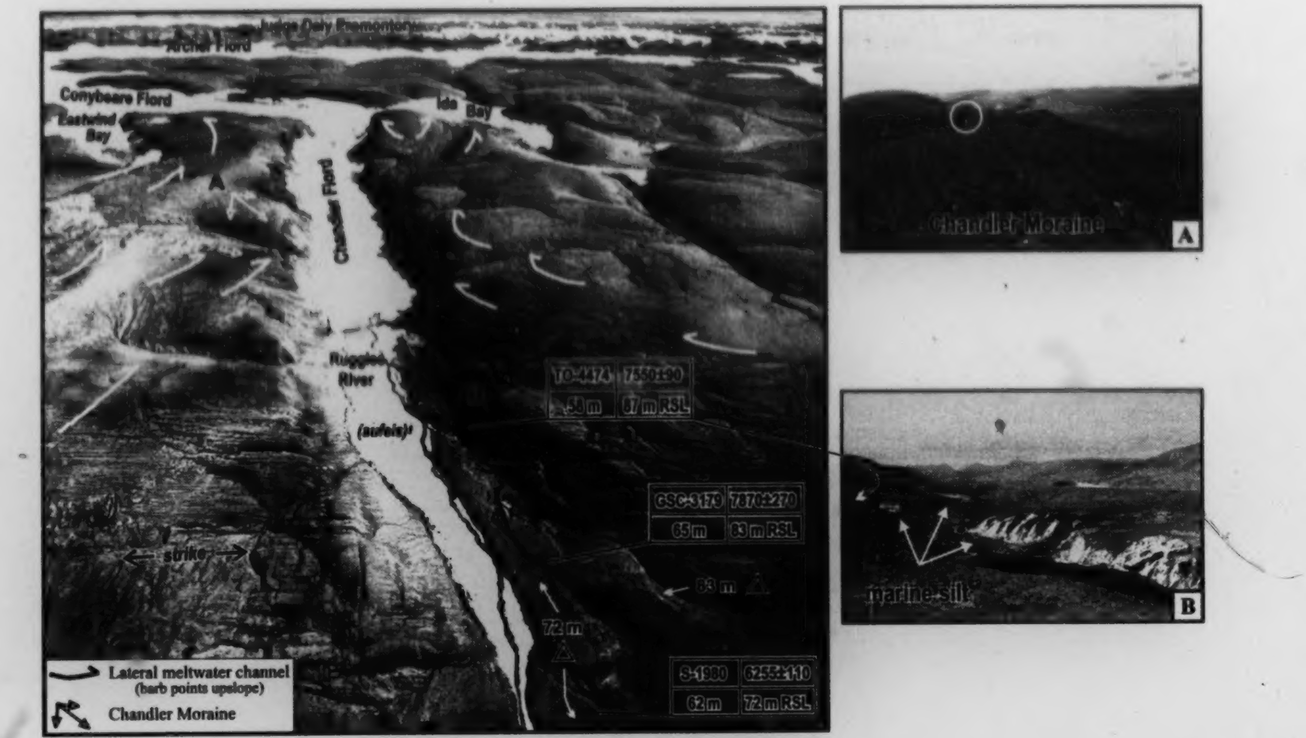

Judge Daly Promontory
Archer Fiord
Conybeare Fiord
Eastwind Bay
Ida Bay
A
Chandler Fiord
Ruggles River
(aufeis)
TO-4474
7550±90
58 m
87 m RSL
GSC-3179
7870±270
65 m
83 m RSL
83 m
72 m
S-1980
6255±110
62 m
72 m RSL
strike
Lateral meltwater channel (barb points upslope)
Chandler Moraine
Chandler Moraine
A
marine silt
B

Figure 5.3 Oblique air photograph looking southeast across the Chandler Fiord region. Marine limit deltas occur along the lower right side of the Ruggles River. Lateral meltwater channels descend out fiord, but do not extend to the plateau summits which are instead covered by felsenmeer. **5.3A** is a close up view of the Chandler Moraine (2 people circled for scale). **5.3B** looks north, up the Ruggles River valley over the southern sections of the 72 m asl delta. Thick sections of pale marine silt outcrop extensively. The 87 m asl terrace occurs at the left center margin of the photo. [Air photograph T490R-118 (©1952) Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada]

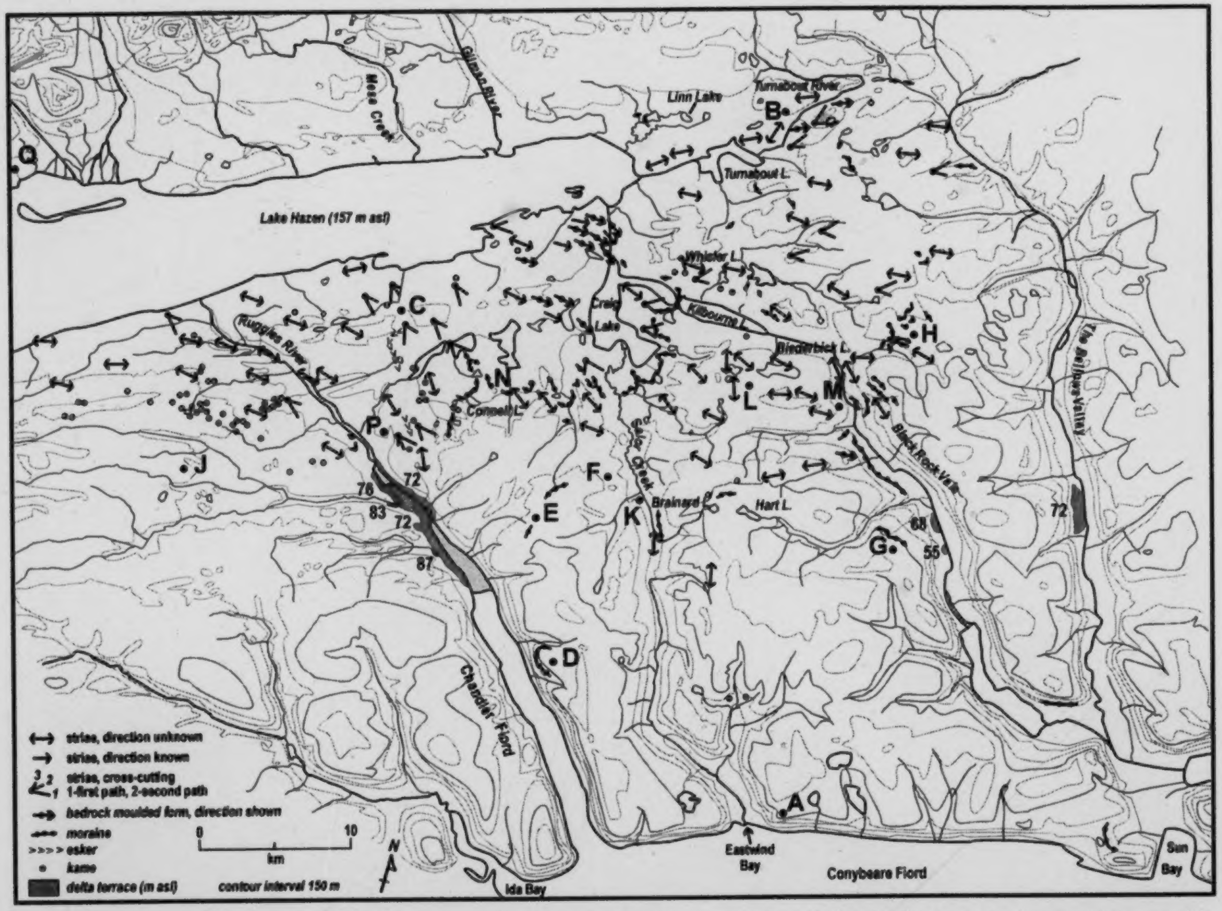

Gilman River
Mesa Creek
Q
Lake Hazen (157 m asl)
Linn Lake
Turnabout River
B
Turnabout L.
Whisler L.
Craig
Lake
Kilbourne L.
Bidderbick L.
C
H
N
Connell L.
L
M
Ruggles River
P
J
F
E
K
Salor Creek
Brainard
Hart L.
Black Rock Vale
The Bellows Valley
G
76
72
83
72
87
68
55
72
D
Chandler Fiord
A
Eastwind Bay
Conybeare Fiord
Sun Bay
Ida Bay
striae, direction unknown
striae, direction known
3 2 1
striae, cross-cutting 1-first path, 2-second path
bedrock moulded form, direction shown
moraine
>>>> esker
kame
delta terrace (m asl)
0
10
km
contour interval 150 m
N

Figure 5.4 Prominent glacial erosional and depositional landforms. Directional indicators (striae and ice moulded bedforms) illustrate a glacial palimpsest in which flow rotated clockwise from northeast to finally southwards out the regional valleys. Bold letters are sites discussed within the text.

margins. Their abundance around the Ruggles River, and relative absence elsewhere in the field area, is related to topographic controls. Around the Ruggles River, ice retreated obliquely, or perpendicular to the south-facing slope, allowing free drainage away from the ice. East of the Ruggles River, retreating ice impounded large proglacial lakes against the rising plateau to the south, depositing instead, deltas and current-bedded sand and gravel.

More than 200 observations of striated bedrock were made throughout the southern field area, the patterns of which demonstrate a glacial palimpsest (Fig. 5.4). South of Lake Hazen, striae were generally well preserved beneath till veneer, but often absent or faint on exposed bedrock. Striae were also absent over most of the upper Hazen Plateau and on Tertiary and Miocene strata north of Lake Hazen. The direction of glacier movement was not always ascertained, although regional patterns of movement are indicated by erosional features including nailhead striae, crescentic gouges, crag-and-tail, roches moutonnées and bedrock flutings.

Grant Land Mountain erratics (notably the chert pebble conglomerate) are found across Hazen Plateau within the core study area (reaching 875 m asl). However, lateral meltwater channels do not extend to summits, which are instead felsenmeer-covered. The boundary between the felsenmeer and uppermost lateral meltwater channel is commonly sharp, and regions below are characterised by relatively clean bedrock, or till veneer (Figs. 5.3 and 5.5). The uppermost meltwater channels along highlands adjacent to Conybeare Fd. occur ~770 m asl, ascending towards the Grant Land Mountains and Agassiz Ice Cap. However, there is considerable local variation in the altitudinal limit of channels, such as around the mouth of Chandler Fd. where they occur below 670 m. Clearly the emplacement of the uppermost erratics requires glacial transport, though it is not implicit that this occurred during the LGM. The absence of meltwater channels on the highest summits and their asymmetry between adjacent highlands suggest that during retreat, "protective" ice caps occupied the summits. The uppermost lateral meltwater channels thus mark the contact with trunk glaciers thinning into the major valleys.

Another distinctive feature of the maximum glacial cover, is the scarcity of glacial depositional and erosional features (notably striae) above 450 m (Fig. 5.4). Arguments by van Tatenhove & Huybrechts (1996), who modelled basal thermal conditions and changing

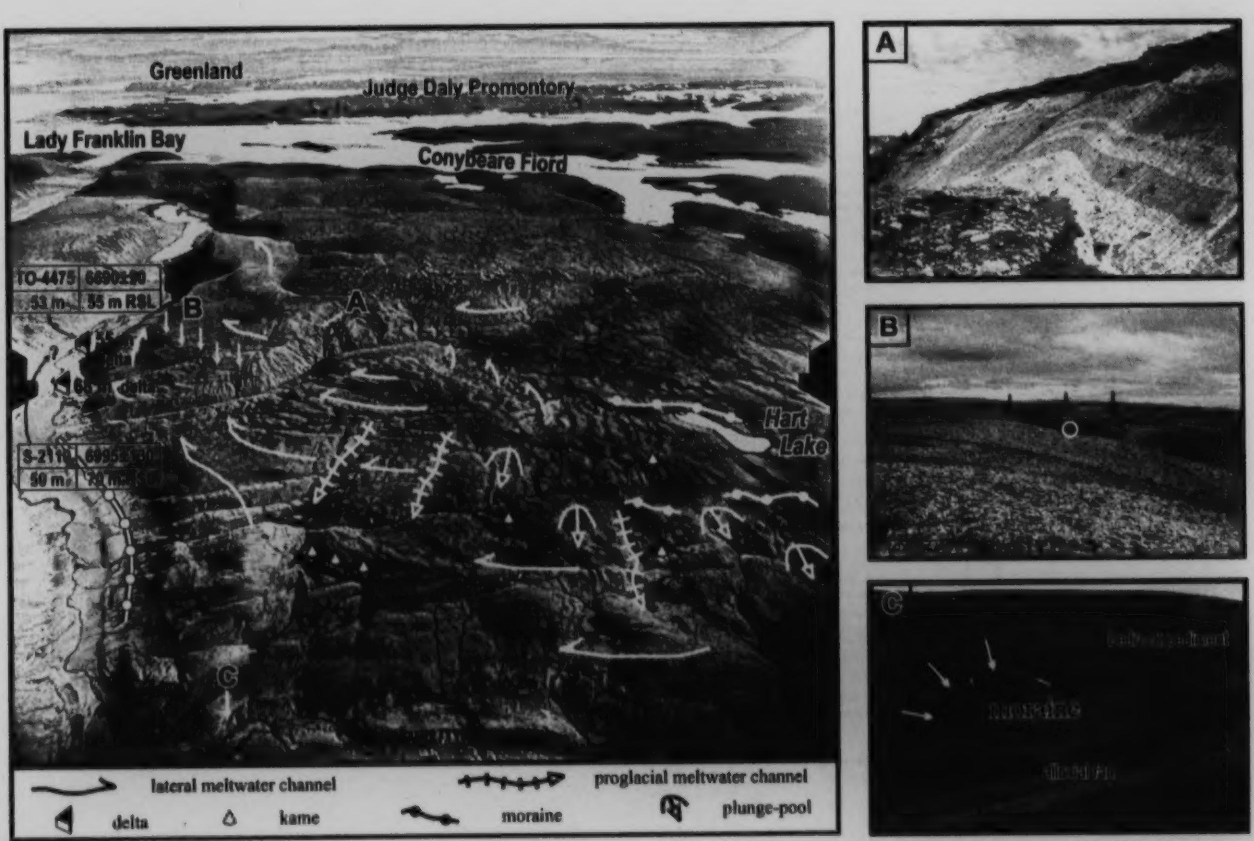

Greenland
Judge Daly Promontory
Lady Franklin Bay
Conybeare Fiord
TO-4475
6690±90
53 m
55 m RSL
B
A
delta
S-2110
50 m
Hart Lake
C
lateral meltwater channel
proglacial meltwater channel
delta
kame
moraine
plunge-pool
A
B
C
moraine

Figure 5.5 Oblique air photograph looking southeast over Hazen Plateau and down Black Rock Vale (on left). Lateral meltwater channels and kames around highlands and the plateau margin record initial retreat of Grant Land Mountain trunk ice. Meltwater channels, traceable upslope to plunge-pools, cross-cut some of the trunk ice lateral meltwater channels, indicating the subsequent expansion of plateau ice caps. **5A** shows a 3 m thick expousre of buried ice (glacial?) containing abundant striated, angular erratics, and capped by a 1 m thick overburden. **5B** looks north along the flat-topped moraine traced by small white arrows on the air photograph. Person circled for scale. **5C** Moraine (white arrows) records a thin outlet glacier retreating from the head of Black Rock Vale. Above this, a flat-lying bedrock pediment of possible Tertiary age unconformably overlies the Danish River Fm. [Air photograph T490R-127 © (1952) Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada]

ice configurations along western Greenland, point to the existence of cold-based outlet glaciers until early Holocene. Their study further models changes in ice geometry, and mean annual ground surface temperature, that lead to temperate bed conditions, and the resultant deposition of larger regional moraines. Based on extensive field observations, Dyke (1993; in press) has similarly argued for the presence of an inner, cold-based zone and an outer, warm-based zone for past ice masses in the Arctic Archipelago. This suggests that if glaciers covered Hazen Plateau during the LGM, they were likely cold-based. Warm-based conditions (and resultant landforms) likely did not form until a later stage of deglaciation when ice became channelled through the major valleys and low-lying areas. Rapid retreat of marine-based margins, may have contributed dynamically to this process.

Additional evidence supporting the hypothesis that Grant Land Mountain ice overrode Hazen Plateau at the LGM, is the presence of what may be buried glacial ice. Exposures of buried ice encircle *Linn Lake* (unofficial name, Fig. 5.4) and outcrop along the distal side of the Craig Lake Moraine (Fig. 5.6, site A). On Hazen Plateau, between Salor Creek and Black Rock Vale, there are several sites where massive ice is exposed, or there is evidence of its occurrence (notably the characteristic slump scars associated with active layer detachments; cf. Harris & Lewkowicz 1993). At one site buried ice, 2-4 m thick, is exposed in a 300 m long headwall (Fig. 5.5, site A). The ice has a foliated appearance, with many foliations becoming truncated at the upper contact with the 0.5-1.5 m thick overburden. Striated Grant Land Mountain and Danish River Fm. erratics are abundant within the ice (Fig. 5.5a), as are discrete lenses of till-like material (several metres in width and up to 1 m thick). The uppermost exposure of ice at this site occurs at 694 m asl. Oxygen isotope analysis ($\delta^{18}O$) of ice samples chipped from the surface of these sites (Table 5.3) support a glacial origin as compositions are well below that of modern-day meteoric waters (~ -28 to -30 ‰; Koerner 1979) and equivalent to those found within the pre-Holocene Agassiz ice core records (Fisher *et al.* 1995).

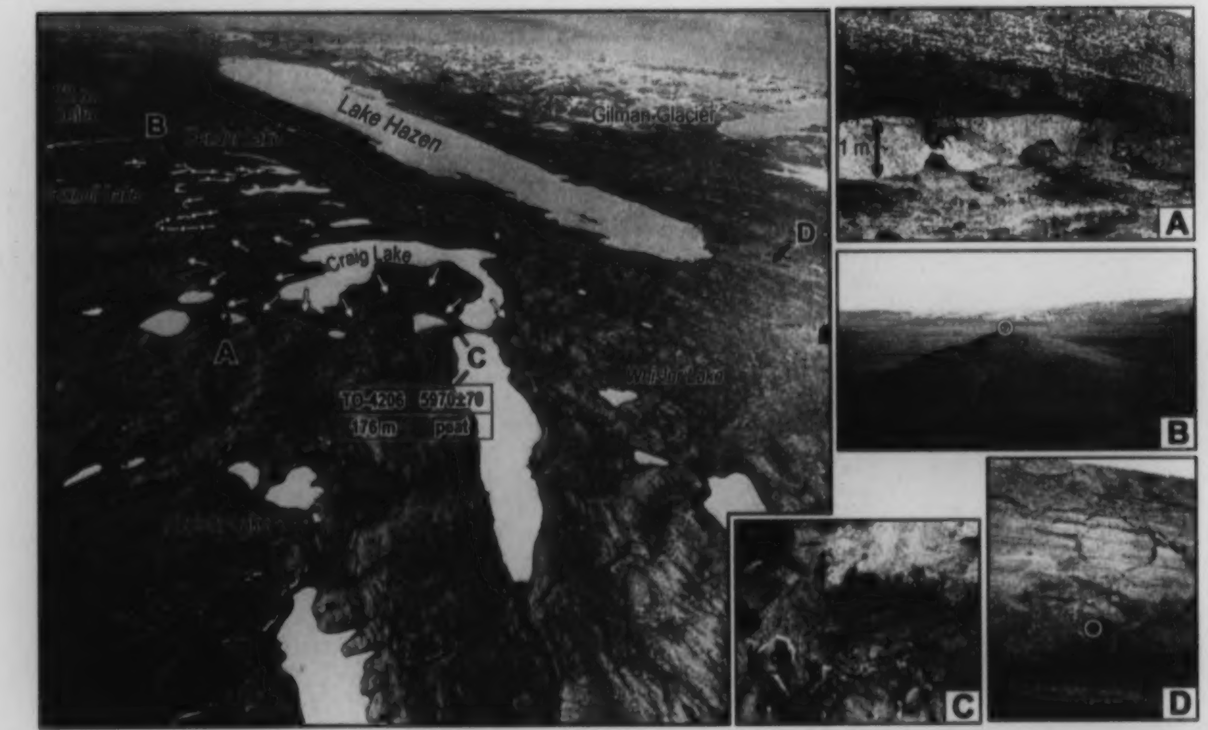


Figure 5.6 Oblique air photograph looking northwest across Lake Hazen Basin. The large, lobate Craig Lake Moraine (white arrows) rings the southern shores of Craig Lake, extending westward as a discontinuous morainic belt. Bold letters show locations of smaller inset photos. **5.6A** buried ice (glacial?) exposed on the distal side of the Craig Lake Moraine; $\delta^{18}O = -37.07$ **5.6B** a large kame in a region of otherwise minimal glacial sediment cover. Person (seated) circled for scale. (photo C. Burnett) **5.6C** Peat underlying a raised delta emanating from the Craig lake Moraine, dated 5970±70 yr BP. **5.6D** Raised, horizontally-bedded lacustrine sediment (~30 m thick). Backpack circled for scale. A lens of organic macrofossils collected 1 m above the base of this section, dated 43 510±690 yr BP. [Air photograph T397L-49 © (1950) Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada]

Table 5.3 Oxygen and deuterium isotope analyses of surface samples of buried ice, exposed within the field area. Results were provided by W. Pollard, McGill University.

Site	$\delta^{18}O$ (‰ SMOW)	$\delta^{2}H$ (‰ SMOW)	Lat. (N)	Long. (W)
Hazen Plateau	-32.50	-245.3	81°45'	67°46'
Craig Lake Moraine (sample 1)	-37.07	-279.1	81°49'	68°37'
Craig Lake Moraine (sample 2)	-37.01	-286.4	81°49'	68°37'
Linn Lake	-34.13	-258.9	81°58'	68°43'

5.5. GLACIAL RECONSTRUCTION

5.5.1 Ice buildup to the Last Glacial Maximum

The growth and coalescence of plateau ice caps were likely amongst the earliest phases of regional glaciation in the study area. Miller *et al.* (1975) indicate that present glaciation levels occur between 1000-900 m asl over much of Hazen Plateau, descending to <700 m asl along the coasts. Much of the plateau lies between 600 - 1300 m asl, and hence several small plateau ice caps exist along the outer plateau rim (Hattersley-Smith & Serson 1972; Bradley & Serreze 1987). Indeed in colder summers (i.e. 1992), snow on the upper plateau surfaces never melts completely. Thus, as the regional equilibrium line altitude (ELA) lowered, widespread glacierization akin to that proposed by Ives *et al.* (1975) on the Baffin Island plateau, may have readily developed over much of Hazen Plateau.

England (in press) suggests that because Greenland erratics are found on outer Hazen Plateau up to 670 m asl, Greenland ice must have had access to this coastline prior to the arrival of Ellesmere ice. However, it may also be that a stronger Greenland ice flow simply displaced pre-existing Ellesmere ice during ice buildup. Once Greenland ice contacted the coast, blocking Robeson Channel, it would have buttressed Ellesmere ice, leading to a rapid thickening. Ellesmere Ice must have quickly inundated the local fiords (water depths between 200 and 600 m) as Greenland erratics are not found above Holocene marine limit within central and inner Lady Franklin Bay (Fig.5.1; England & Bradley 1978). This suggests that when Greenland ice was advancing onto outer Hazen Plateau, Grant Land Mountain ice had already infilled Lake Hazen Basin. Expansion of Ellesmere ice prior to this was likely limited by an inability to overcome fiord depths.

The absence of evidence for flow of Agassiz ice inland of the head of Chandler Fiord (i.e., erratics, northward descending meltwater channels), suggests a congruent buildup of Agassiz and Grant Land Mountain ice at the head of Conybeare Fiord. Distinctive erratics, including a fossiliferous mudstone and sandstone (Fig. 5.7), found within 300 m of the plateau edge above Eastwind Bay (~750 m; Fig. 5.4, site A), do, however, suggest that ice flow from the southwest (Agassiz source) eventually overtopped the outer rim of Hazen Plateau. Where these erratics outcrop is unknown. However, as they were not found anywhere else in the field area, it implies they outcrop west of here, and have been carried

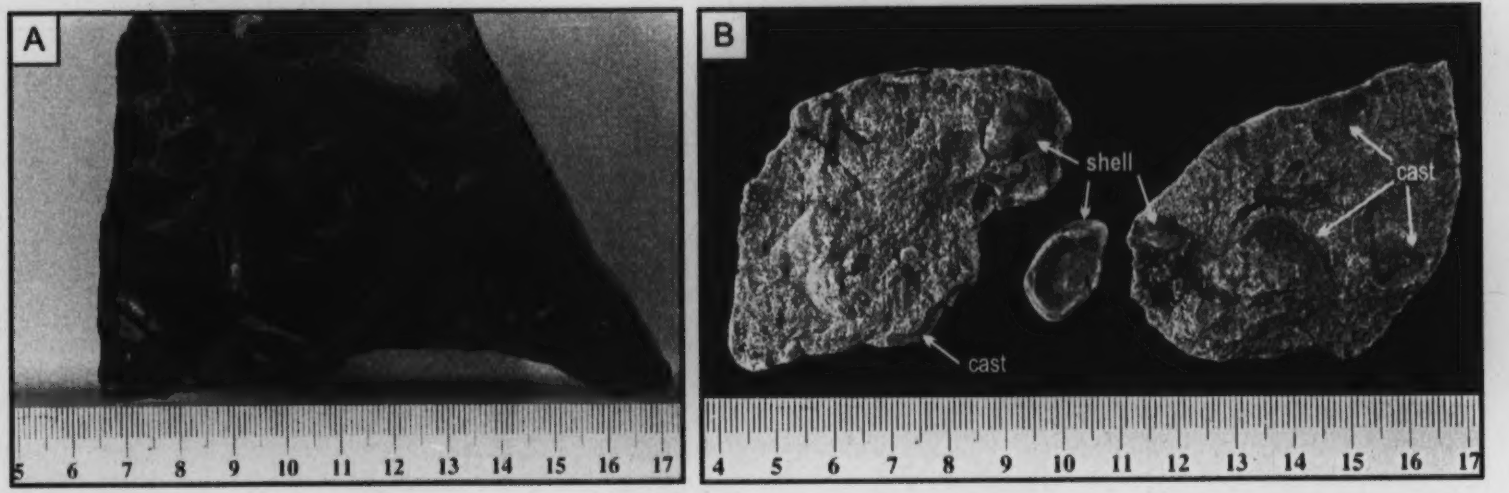

A
5 6 7 8 9 10 11 12 13 14 15 16 17
B
shell
cast
cast
4 5 6 7 8 9 10 11 12 13 14 15 16 17

Figure 5.7 Fossiliferous erratics found above Eastwind Bay (750 m asl). Scales in centimeters. The outcrops of these erratics are unknown, but are assumed to lie west of the field area, towards the Agassiz Ice Cap. **5.7A** single valves of an unknown bivalve, 9-24 mm long, exposed along a single lens within a massive mudstone. **5.7B** single valves of an unknown bivalve, 9-11 mm long, found throughout a medium-grained sandstone.

by Agassiz ice flowing northeast down the Dodge River valley. During the emplacement of the erratics, flow of Grant Land Mountain trunk ice southward out the fiords would have been blocked, resulting in a deflection inland. This is recorded by striae along the Turnabout River (Fig. 5.4, site B), which do not trend towards the outlet valleys leading to Lady Franklin Bay, but instead trend northeastward towards the Grant Ice Cap. During this phase, the trunk ice would have deflected the Gilman and Turnabout glaciers along the south slope of the Grant Land Mountains. Piper Pass (~274 m asl) would have served as a conduit, channelling ice northwards into Clements Markham Inlet.

5.5.2 Initial Deglaciation of Hazen Plateau

A second set of striae located south of Lake Hazen, indicate a rotation of flow from northeast (inland) to south, gradually becoming parallel to the orientation of the major valleys and fiords. While striae found along the valleys may record topographic control, several striae found atop the ridge immediately south of Lake Hazen (Fig. 5.4, site C) record a flow perpendicular to the underlying structure and Lake Hazen trough, indicating a flow unhindered by topography. North of Lake Hazen, lateral meltwater channels along the North and South Boulder Hills (Fig. 5.2) continue to record a northeastward flow across Hazen Plateau.

Thinning of the pervasive ice cover and separation between outlet glaciers and remnant plateau ice caps is recorded by low gradient (6-10 m/km), lateral meltwater channels ringing local plateau uplands, descending southeastward towards the coast (Figs. 5.3 and 5.5). The profile of the highest meltwater channels bordering the coast necessitates the presence of ice infilling Lady Franklin Bay and Robeson Channel. Further, if Greenland ice coalesced with Ellesmere ice along Hazen Plateau as suggested by England (1998, in press), then these same outer meltwater channels would have formed only after the coalescent boundary lay either along the coast, or offshore. The pattern of channels observed, also records the progressive thinning of ice from the northeast, southwards towards the Agassiz Ice Cap and northwards towards the Grant Land Mountains.

Retreat is further documented by the prominent Chandler Moraine, which flanks the east side of Chandler Fiord (outer Hazen Plateau) at 622 m asl (Fig. 5.3; Fig. 5.4, site D).

Its gradient records a southwards ice flow that was not fed by local plateau ice, because the moraine descends, rather than rises, into the adjacent side valley.

An estimate of the age of the moraine is provided by cosmogenic ^{36}Cl dating. Cosmogenic isotopes, such as ^{36}Cl, accumulate in near-surface materials such as soil, rock and ice, the measurement of which has been used to estimate exposure-age, weathering and surface stability (Briner & Swanson 1998; Bierman 1994; Hallet & Putkonen 1994). Dates are, however, artifacts of the models used to interpret isotope abundance, and are subject to many uncertainties with respect to latitude, altitude and atmospherics (Conrad *et al.* 1989; Bierman 1994). For the most part, ages should be considered maximum estimates, as failure to remove all pre-exposed material (^{36}Cl production occurs beyond 3 m depth; Bierman 1994) would lead to a date that is too old. Environmental factors such as weathering (which removes outer exposed surfaces), extensive snow cover (which precludes cosmic rays below 2 m water equivalent, M. Zreda, pers. comm. 1996), and gelifluction which may alter the position or exposure can all lead to an incorrectly lower age estimate. It is thus preferable to analyse several samples at each site, giving preference to the youngest ages. Only erratics that retained striae (indicating minimal post-depositional weathering), and bedrock in areas that are unlikely to retain prolonged seasonal snow cover were sampled.

Cosmogenic ^{36}Cl dating of striated erratic boulders on the Chandler Moraine yielded an average age of 18.5±3.1 ka (sidereal years; Table 2; M. Zreda, pers. comm. 1996). *En echelon* lateral meltwater channels above and below the moraine, indicate that it marks a recessional position. The date should be considered a maximum estimate, but clearly assigns the moraine to the late-Wisconsinan. This corroborates several ^{36}Cl dates of late-Wisconsinan age obtained on bedrock and erratics along eastern Ellesmere Island (M. Zreda, pers. comm. 1998).

Moraines along the north-facing slope of the plateau west of Salor Creek (~520 m asl; Fig. 5.4, site E), and lateral meltwater channels wrapping around the plateau, record subsequent deglacial margins and southeastward flow. Meltwater channels below this site extend to a delta in a side valley of Salor Creek at 362 m asl (Fig. 5.4, site F), indicating the impoundment of an ice marginal lake against trunk ice flowing southwards up Salor Creek.

Near the mouth of Black Rock Vale, the uppermost lateral meltwater channels (770 m asl) are low gradient (~6 m/km), increasing to 26 m/km around 680 m asl, then returning to low gradients (~6 m/km) around 300 m asl (Fig. 5.8). Their configuration is likely controlled by the dynamics of ice infilling the marine margins, suggesting that during the intermediary phase, there was a rapid retreat. Channelling of flow within the confining valleys contributed to an over-steepened profile as evidenced by lateral meltwater channels feeding a meltwater canyon at 660 m asl, which descends abruptly to a prominent lateral meltwater channel parallelling the coast at 530 m (indicating that much of the outer plateau rim was already deglaciated; Figs. 5.8 and 5.9). The gradient on the lateral meltwater channel starting at 530 m is ~14 m/km, which if projected northwards, would reach local marine limit (100-116 m asl), just beyond the mouth of Lady Franklin Bay (albeit this would represent a minimum estimate as ice surface gradients are likely to shallow towards the margin). The importance of this is that it suggests that locally, deglaciation of much of the upper plateau may not have occurred until after initial marine limit deltas formed along northern Robeson Channel. Estimates of this are provided by dates of: i) 8600±90 BP on *in situ* shells at 91 m in Lincoln Bay (immediately north of Wrangel Bay) which is a minimum estimate for local marine limit (~100 m; Retelle 1986), and ii) 9580±140 BP on paired shells in marine silts at 110 m (relative sea level <140 m) from central Hall Land, Greenland (England 1985).

Along the north-facing slope of the plateau north of *Brainard* and *Hart* lakes (unofficial names), kames and small moraines, between 620 and 570 m asl, record ice wrapping around the plateau, flowing from west to east (Figs. 5.4, 5.5 and 5.9). Down-valley from this, a prominent moraine is found along west-central Black Rock Vale (Fig. 5.4, site G; Fig. 5.5, site B) at 541 m asl. It has a broad (25 m), flat surface profile (Fig. 5.5B), and descends into a side valley (Fig. 5.5), indicating that trunk ice in Black Rock Vale did not coalesce with plateau ice flowing out of this side valley.

Northeastward across the valley from Biederbick Lake, a small (5 m high), sharp-crested moraine occurs at 515 m asl (Fig. 5.4, site H), and extends eastward along the crest of a ridge. The configuration of this and adjacent moraines nested below, record flow of the Grant Land Mountain trunk glacier southeastward against the plateau, descending into the side valley. Cosmogenic ^{36}Cl dating of a striated boulder from this moraine provided an age

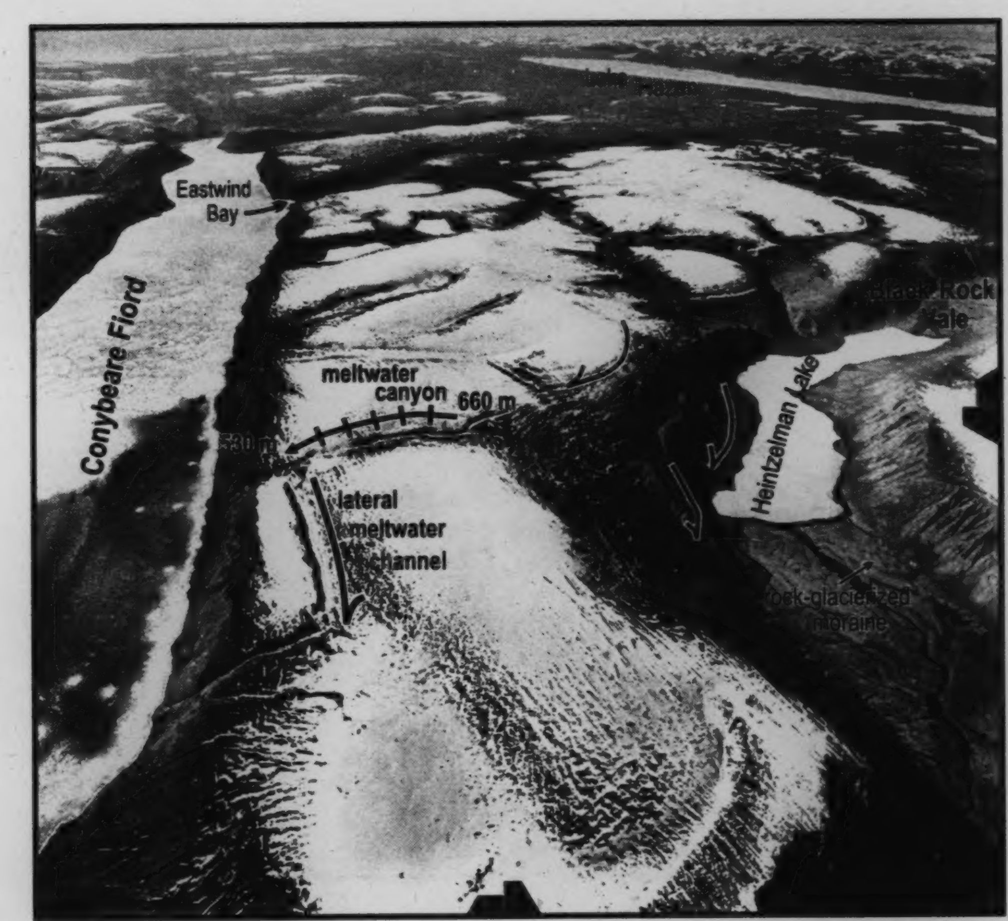

Eastwind
Bay
Conybeare Fiord
meltwater
canyon
660 m
530 m
lateral
meltwater
channel
Black Rock
Vale
Heintzelman Lake
rock-glacierized
moraine

Figure 5.8 Oblique air photograph looking west up Conybeare Fiord and Black Rock Vale. When trunk ice occupied Black Rock Vale it fed a large meltwater canyon (660 m asl) that extended southwards to a prominent lateral meltwater channel (530 m asl) impounded by ice within Conybeare Fiord. Projecting the gradient of this channel downflow (northeast) indicates that it would reach local marine limit (~116 m asl; ~9 ka BP) beyond Lady Franklin Bay. Hence, the transition of lateral meltwater channel gradients within Black Rock Vale from 6-10 m/km (~770 m asl) to 26 m/km (~680 m asl), then back to 6-10 m/km (below 300 m asl) may record the rapid retreat of marine-based ice between 9 and 8 ka BP, at which point Grant Land Mountain trunk ice became grounded in shallow fiords. [Air photograph T397R-138 © (1950) Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada]

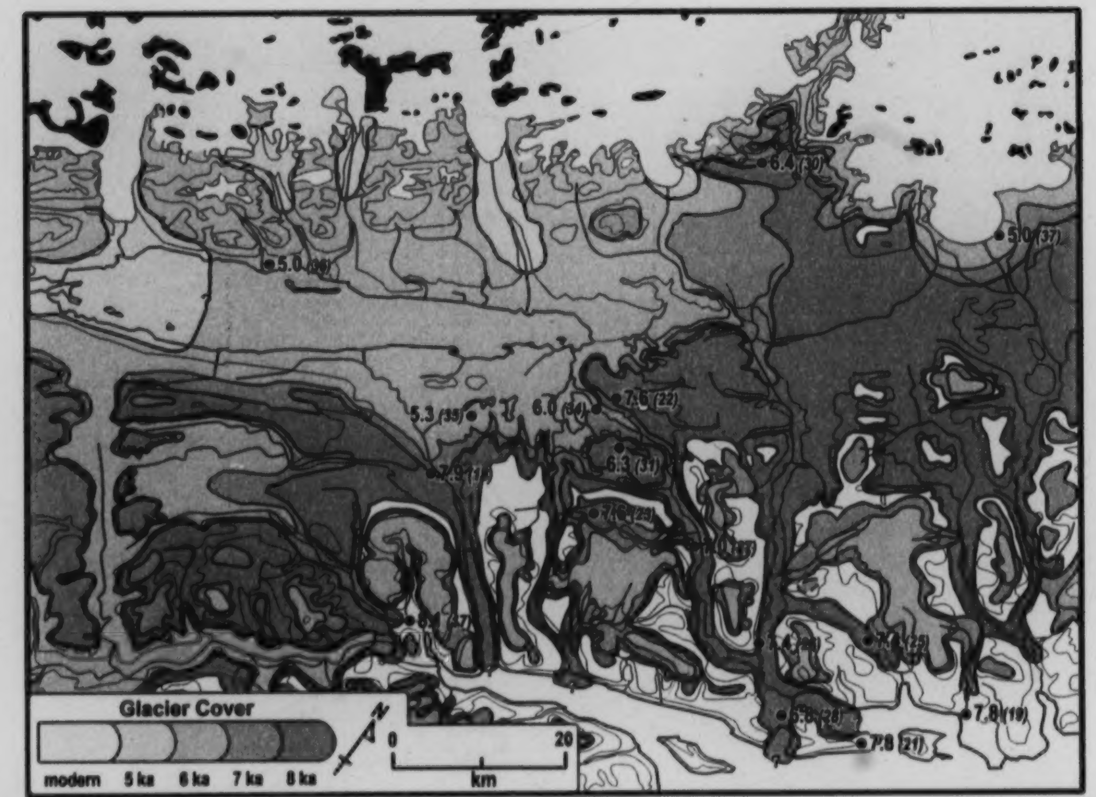


Figure 5.9 Paleogeographic map depicting changes in ice margins at 8, 7, 6 and 5 ka BP. Key radiocarbon dates are indicated in thousands of years before present (ka BP), site number from Table 1 shown in brackets. Contour interval 300 m. Former margins of the Grant Land Mountain trunk ice are generally well defined by lateral meltwater channels, moraines and ice-contact deltas. The margins of plateau ice caps, however, are generally poorly constrained, and therefore should be considered speculative. The reconstructions illustrate the transition from complete ice cover (~10 ka BP), through the retreat and thinning of trunk ice within the regional fiords and valleys, to the final breakup within Lake Hazen Basin (~5 ka BP). Plateau ice caps expanded in the early Holocene (cross-cutting former lateral meltwater channels of the trunk glacier), and then persisted until the mid-Holocene.

of 8±0.5 ka (Table 5.2). By comparison, dates on glacially polished bedrock west of the Ruggles River yielded an average age of 14.3±4.1 ka (Fig. 5.4, site J; Table 5.2). Given the elevation of the Ruggles River sample (310 m), and its position with respect to the source and direction of trunk glacier flow, it could not have been deglaciated prior to the establishment of the moraine north of Black Rock Vale (8 ka). Hence it is likely that the date of 14 ka records some pre-exposure not removed by glacial abrasion. Furthermore, a striated boulder on the Craig Lake Moraine dated 9.8±1.5 ka. This moraine also was deposited after the 8 ka moraine downvalley, and is constrained by a peat date of 5970±70 BP (Tables 5.1 and 5.2). The older date from the Craig Lake Moraine must also indicate pre-exposure of the boulder. The inability to correct for this indicates that the utility of ^{36}Cl to resolve late-Wisconsinan (Holocene) deglacial sequences is questionable without large numbers of samples.

5.5.3 Marine Incursion

Ice margins that advanced beyond Hazen Plateau entered Lady Franklin Bay where they were marine-based. Retreat of ice from Lady Franklin Bay is constrained by dates of 8520±80 BP from midway up Archer Fiord, 8380±105 BP from shells in the topset beds of the Cape Baird Terrace, and 8130±200 BP in Ida Bay at the head of Conybeare Fd.(Fig. 5.9; Table 5.1). Younger marine limit ages recorded outside these sites (England 1978; 1983) are best considered minimum estimates, or point to the continued occupation of valleys by outlet glaciers. Marine limit at the head of Chandler Fd. has been reported to be 83 m asl, marked by a prominent gravel terrace, overlying proximal bottomset silt. *In situ Portlandia arctica* from the silt dated 7860±270 BP (Lemmen 1981; England 1983; Table 5.1). However, 5 km down-fiord from this site, thick marine silt (extending up to 58 m asl), underlies a washed gravel surface at 87 m, extending from a side valley which would have drained ice inland of the head of Chandler Fiord and local plateau ice caps (Fig. 5.3). The marine silt contains abundant, mostly single valves of *Hiatella arctica*, *Mya truncata* and *P. arctica*. A paired valve of *P. arctica* collected from the top of these silts dated 7550±90 BP (Table 5.1). This now establishes a higher marine limit (≥87 m), but remains constrained by the greater age of 7860±270 BP, which itself remains a minimum estimate. A massive silt deposit (barren of shells) at 91 m asl, 2 km north of the 87 m terrace may provide a better minimum elevation for marine limit. Lateral meltwater channels traced from Lake Hazen Basin beyond the head of Chandler Fiord, descend to ~100 m asl, and therefore represent a maximum estimate for marine limit.

Two prominent ice-contact deltas are found midway up Black Rock Vale, at 68 and 55 m asl (Figs. 5.4 and 5.5). The age of the uppermost delta is constrained by a date of 6995±130 BP (Table 5.1) on shells from marine silt at 50 m, related to a relative sea level of ~70 m asl (England 1983). Paired *P. arctica* valves from the sandy topsets of the 55 m delta dated 6690±90 BP (Table 5.1). Marine limits along northern Lady Franklin Bay range from 111 to 115 m (England & Bednarski 1986). Hence, the absence of marine deposits above the 68 m delta indicates that the valley below the deltas must have been blocked by ice prior to their establishment. A large moraine running along the valley bottom of upper Black Rock Vale (Figs. 5.4 and 5.5) and what are interpreted to be large, rock glacierized moraines at the confluence with The Bellows Valley (Fig. 5.8), argue that Grant Land Mountain ice was likely the dominant contributor to ice occupying the valley. However, a confluent flow of ice from the adjacent western valley feeding off the plateau also appears likely, as the large moraine terminates abruptly at the entry of this valley (Figs. 5.4 and 5.5). Also, subsequent to the establishment of the 68 m delta, a large boulder fan prograded from this side valley, indicating an abundant meltwater source from the plateau.

Marine limit within The Bellows Valley is recorded by a small raised delta (115 m asl) near the mouth of the valley, which is undated (England, pers. comm. 1998). Minimum estimates are provided by shell dates of 7755±125 BP from a surface collection (102-105 m) at Sun Cape, and 7385±375 BP collected in bottomset sands (32 m asl) just up-valley from this site (Fig. 5.9, Table 5.1; England 1983). Below the 115 m delta, no other deltas occur until an extensive delta was deposited at 72 m asl, 25 km up-valley (England, pers. comm 1998; Fig. 5.4). This feature remains undated, but must predate the 68 m delta in Black Rock Vale (7 ka). What is remarkable about this delta is its extent, the abundance of Tertiary wood within it (England 1983) and the volume of sand and gravel that has infilled the valley (45 km long). It is proposed that this conspicuous sedimentation corresponds to the drainage of a large proglacial lake that formed at the head of The Bellows Valley (Fig. 5.2B). The former proglacial lake is recorded by extensive blankets of lacustrine sediment, and a descending sequence of deeply incised spillways marking former outlets that progressed westward with the retreating ice margin (Fig. 5.2B). The date of 7385±375 BP (Table 5.1) within the bottomset sand at the head of the valley may provide a minimum estimate on the drainage of this lake, and has been suggested by England (1983) to constrain the 72 m delta up-valley. Slow retreat of land-based ice from Black Rock Vale and The Bellows Valley is therefore recorded by the drop in marine limit from 115 m asl (7.8 ka) at the coast to 68 m asl (6.8 ka), 25 km inland (Fig. 5.9).

5.5.4 Plateau Ice Caps

There are two sets of meltwater channels along the plateau north of Hart Lake (Fig. 5.5). The first of these is a series of lateral meltwater channels that encircle and descend from the highland summit (759 m asl) adjacent to Hart Lake, down to the valley bottom, wrapping west to east around the plateau and southwards down Black Rock Vale (Fig. 5.5). These record the retreat of Grant Land Mountain trunk ice. The second set of meltwater channels is oriented perpendicular downslope from the plateau north of Hart Lake. These channels can be traced upwards to conspicuous plunge-pools along the plateau margin (~600 m asl; Fig. 5.5), likely indicating a supraglacial to proglacial descent of meltwater. They cross-cut trunk ice lateral meltwater channels, but do not extend to the base of the slope, indicating that trunk ice continued to occupy the valleys below (Fig. 5.5). Furthermore, lateral meltwater channels relating to the thinning of the trunk ice are also found above the plunge pools, indicating that local plateau ice caps expanded during the early Holocene.

With a few exceptions, it is difficult to constrain the age and extent of the plateau ice caps. A date of 7550±80 BP on a mat of bryophytes immediately overlying a diamict in the Brainard Lake core (Table 5.1) provides a minimum estimate of plateau ice retreat. This same plateau ice cap (between Salor Creek and Black Rock Vale) must have persisted until mid-Holocene because an outlet glacier flowing eastward from it can be linked to the ice-contact deltas (6995±130 and 6690±90 BP; Table 5.1) and a large alluvial fan in Black Rock Vale. It also dispersed an outlet glacier that impounded a lake between it and the northward retreating Grant Land Mountain ice within Salor Creek valley, indicated by raised deltas and lacustrine sediment (Figs. 5.9 and 5.4, site K). The age of this proglacial lake is uncertain, but would predate the formation of the Craig Lake Moraine (~6 ka) based on the regional glacial geomorphology. Persistence of this, and a plateau ice cap west of Salor Creek, is demonstrated by the extensive raised delta that prograded northwards from Salor Creek into Craig Lake, at a time when Grant Land Mountain ice had retreated from the Craig Lake Moraine (~6 ka), but continued to block the modern outlet at the north end of Craig Lake.

5.5.5 Retreat of Ice Inland From Hazen Plateau

As the Grant Land Mountain trunk ice continued to retreat, its flow path shifted in response to increasing topographic control. Abundant striae record the trunk ice advancing northeast through the Lake Hazen trough, spilling obliquely across the ridge south of Lake Hazen, and then being drawn down the major valleys (Fig. 5.4). Striae and bedrock flutings (1-3 m high, 10-40 m long) along the southeast end of Lake Hazen record a strong deflection of flow, likely between the trunk glacier, and the coalescent Gilman Glacier (Fig. 5.4). The shifting of the coalescent boundary, is recorded by the progressive deflection of flow from northeast to southeast, indicating increased draw-down into Black Rock Vale and The Bellows Valley. Over the Appleby/Biederbick Lake upland (Fig. 5.4, site L), striae conform to small topographic features.

Subsequent to the establishment of the marine limit delta at the mouth of the Ruggles River (7.9 ka BP), ice retreated less than 5 km up valley, before it formed the extensive 72 m delta, dated 6100±260 BP (Table 5.1). Elsewhere, ice retreated much more extensively during the same period (Fig. 5.9).

Grant Land Mountain trunk ice retreated northwards and westwards from the mouth of the Ruggles River exposing a large valley west of the Ruggles River (Fig. 5.9 and 5.4, site J). Kames and till within this region have unusually thick carbonate pendants on the underside of clasts (often encasing the entire matrix) compared to those found immediately north of here. This, and the presence of rare erratics found on deeply weathered bedrock pedestals, suggest considerably greater antiquity for this site. However, this is not supported by the regional glacial reconstruction, and indeed, neither bedrock weathering, nor the presence of tors, could be accurately used to demarcate relative ages on the carbonate-dominated terrain throughout the field area. In fact, some of the most well developed tafoni occur on bedrock and boulders inside the 6 ka BP ice margin.

A single fragment of the bryophyte *Drepanocladus brevifolius*, found within the upper diamict of *Whisler Lake* (unofficial name) sediment core, dated 7600±400 BP (Table 5.1). Arguments presented in Chapter 3 of this thesis suggest that the diamicts found in cores of this and other lakes are of non-glacial origin. Interpretation of the 7.6 ka BP date, however, is uncertain, as the bryophyte is redeposited, and therefore may have been transported from deglaciated sites in the Grant Land Mountains, or the local catchment, requiring ice to have thinned below ~300 m. This remains a possibility, given the pattern of lateral meltwater channels connecting with the marine limit delta (7.9 ka BP) at the head of Chandler Fiord, which point to ice margins below 300 m along southern Lake Hazen Basin (Fig. 5.9). Continued thinning and retreat of ice led to a channelling of flow within the Kilbourne Lake valley, recorded by a series of kame terraces along the valley walls (Fig. 5.4). A proglacial lake also formed within this valley, impounded by a till plug at the head of Black Rock Vale (Fig. 5.4, site M). Breaching of this barrier, and successive stages of the lake are recorded by staircases of deltas deposited along streams entering the basin, including large raised deltas, 11 m above the modern stage, emanating from the Craig Lake Moraine (~6 ka BP).

Towards Piper Pass, LaFarge-England *et al.* (1991) report a date of 6410±50 BP from peat at 450 m (actual base of the peat was undetermined), providing a minimum estimate for deglaciation (Fig. 5.9). The proximity of this site to the Turnabout and other glaciers (<10 km), points to a substantially reduced ice cover. However, these glaciers are fed by a relatively small catchment, separated from the ice cap feeding the trunk glacier that continued to occupy Lake Hazen Basin.

Further retreat led to the establishment of the Craig Lake Moraine, the largest and most extensive moraine system in the study area (Figs. 5.4 and 5.6). It is highly lobate, possibly the product of surging, and its size (up to 43 m high) is enhanced by the presence of buried ice (glacial), exposed on its distal side (Fig. 5.6, site A; Table 5.3). The age of this moraine is constrained by a date of 5970±70 BP (Table 5.1) from the base of 1.5 m of peat, underlying a proglacial delta emanating from the moraine (Fig. 5.6, site C). A discontinuous morainic belt extends ~25 km westward from the Craig Lake Moraine, and can be traced to the ice-contact 72 m delta along the Ruggles River (Figs. 5.3, 5.4 and 5.6). Dates on paired, *in situ* shells from the foreset beds of the 72 m delta, range from 6255±110 to 6100±260 BP (Table 1), supporting its coeval formation. One section of the adjoining morainic belt, east of *Connell Lake* (unofficial name; Fig. 5.4, site N), contains abundant klinker, a unique erratic, likely derived from coal beds exposed along the north shore of Lake Hazen and rivers dissecting the Tertiary strata (Miall 1979). Northwest of Connell Lake, an anastomosing esker (5-20 m high; one of only two large eskers observed within the field area) originates at a low pass in the ridge, then makes a 90° bend downslope, terminating at the bottom of the slope. The anastomosing pattern indicates open conduit flow (cf. Hooke 1984), under clearly warm-based conditions. Smaller (<2 m high), discontinuous eskers can be traced southward from the main esker towards the 6 ka BP ice margin. East of the Craig Lake Moraine, the 6 ka BP ice margins are provisional, based on undated lateral meltwater channels and other geomorphic evidence (Fig. 5.9).

A date of 5290±110 BP (Table 5.1) was determined on bryophytes from a lens of sorted sand and gravel within a core from Connell Lake. These sediments must be ice-proximal because there is no drainage into the basin today which is capable of transporting gravel. Outlet channels and small deltas marking this earlier proglacial stage were active when a tongue of ice still occupied the central Ruggles River. However, it is uncertain if the trunk glacier continued to overtop the ridge south of Lake Hazen, or whether the sediment source was from a residual ice mass within the basin.

Multiple, cross-cutting striae in the region immediately east of the Ruggles River, record the final phases of flow, that was first oblique across the Ruggles River valley, then parallel, and finally, convergent into the valley (Fig. 5.4, site P). Absence of similar radiate patterns around Craig Lake indicates the separation of trunk ice west of Craig Lake along a topographic high. Along the south shore of Lake Hazen, striae remained parallel with the axis of the trough.

Retreat of ice from Lake Hazen proper is constrained by a minimum date of 4980±70 BP (Table 5.1) on a peat deposit exposed 1.5 km north of Lake Hazen on Skeleton Creek (230 m; Fig. 5.4, site Q; Gould *in* Blake 1986). Similarly, a date of 4970±60 BP on peat found 60 m beyond the modern margin of the Eugene Glacier (Lowdon & Blake 1979) confirms that by ~5 ka, Grant Land Mountain ice had retreated to near its modern limits throughout the study area. Thus, melting of residual ice within the interior of the Lake Hazen region cannot be linked to a late mid-Holocene sea level rise (cf. Scott & Collins 1996).

The final, rapid retreat of ice in Lake Hazen Basin was likely accentuated by calving within proto-Lake Hazen. Evidence of proto-Lake Hazen and the sequential westward retreat of ice within the basin is documented by numerous deltas (204-169 m asl) and current-bedded sediments around the eastern end of Lake Hazen, including the Turnabout River, where an eastward flow was opposite to present. Extensive dissected lacustrine sediments at the east end of Lake Hazen (Fig. 5.6D), exposures of which range up to 32 m thick, record former lake levels of 204, 189 and 169 m asl (currently Lake Hazen is 157 m). A lens of plant macrofossils exposed at the base of the section shown in Figure 5.6D, contains large pieces of coalified wood which were rejected for dating. Instead, "clean" and intact *Hippuris vulgaris* and *Potamogetan* spp. seeds (aquatic plants) were separated under a microscope and dated 43 510±690 BP (Table 5.1). Several other old dates (>30 ka) have been obtained around Lake Hazen. These include proglacial lake sediments along Mesa Creek, whose sand/silt rhythmites alternated with lenses of *Cassiope tetragona* that dated 37 260±470 BP (Table 5.1), indicating the reworking of pre-glacial organic material. The lower proto-Lake Hazen level recorded here (169 m) can be traced around much of the eastern and central shore today, and records the final blocking of the modern Ruggles River outflow. A difference of 2-3 m between this raised shoreline along the south shore (~171 m) and its counterpart on the north shore (~168 m) indicates differential postglacial emergence towards the Grant Land Mountains, which is consistent with regional isobases plotted by England & Bednarski (1986).

5.6 DISCUSSION AND CONCLUSIONS

5.6.1 Last Glacial Maximum

The last glaciation of Lake Hazen Basin and eastern Hazen Plateau was characterized by a pervasive Grant Land Mountain ice cover, which coalesced with Agassiz ice flowing northeastward out of Conybeare and Archer fiords (Fig. 5.10a). Extensive biological refugia did not therefore exist over Hazen Plateau, although it remains to demonstrated whether they existed on exposed sections of the continental shelf, or on nunataks along northern Ellesmere Island and Greenland.

The growth of ice up to the LGM remains poorly understood both on Hazen Plateau and throughout the Arctic Archipelago (Dyke, in press; England, in press). Evidence from the Agassiz Ice Cap suggests that ice buildup following the last interglacial may not have occurred until middle to late-Wisconsinan (Koerner *et al.* 1987). A similar reconstruction has been proposed for northwestern Greenland, in which the Independence Fiord glaciation (LGM) is suggested to have begun around 40 ka, attaining its maximum extent at, or before, 14 ka BP (Funder 1989; Funder & Hansen 1996). On Ellesmere Island, a maximum age for ice advance is provided by dates on individual shells in tills along the east-central coast as young as 19 ka BP, and on the northeast coast (and the adjoining Hall Land, Greenland) that date 23-24 ka BP (England 1998, in press). Similar dates have been reported on northern Ellesmere Island (Lemmen 1989; Evans 1990), including a date of 23 850±850 BP (Table 5.1) measured on a *Salix arctica* twig redeposited in deglacial sand at the head of Clements Markham Inlet (Bednarski 1986).

Within the study area, striae along the Turnabout River and unique erratics above Eastwind Bay record an inland deflection of ice flow. This thus confirms the presence of coalescent Ellesmere and Greenland ice along outer Hazen Plateau during the last glaciation (England 1998, in press), but does not address the exact position and dynamics of this boundary. Future field investigations designed to trace the trajectory of the erratics identified above Eastwind Bay may help to resolve this. England (1998, in press) has proposed that the coalescent boundary lay along the western limit of granite (Greenland) erratics (Fig. 5.10B). While their age is implicitly ascribed to the LGM (England 1998, in press), they remain to be dated by cosmogenic means. Evidence from cores from the Agassiz Ice Cap (Koerner *et*

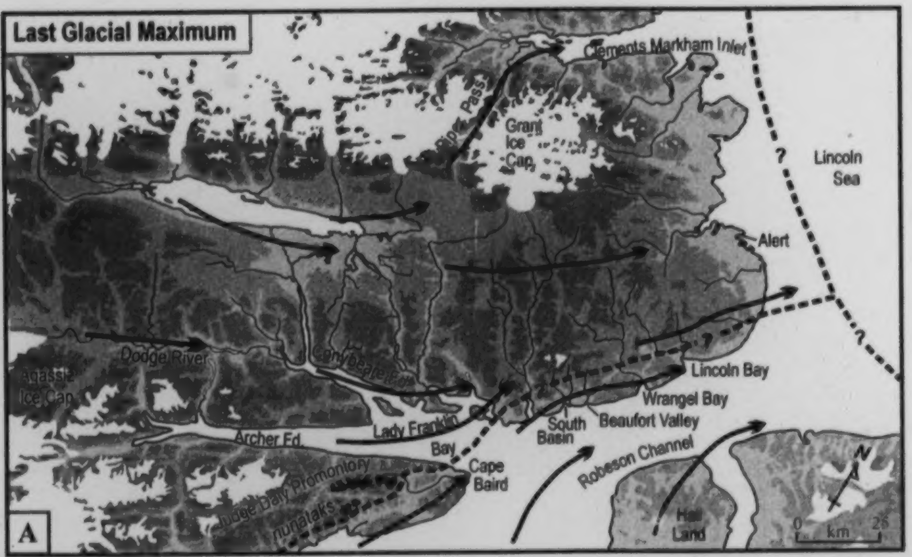
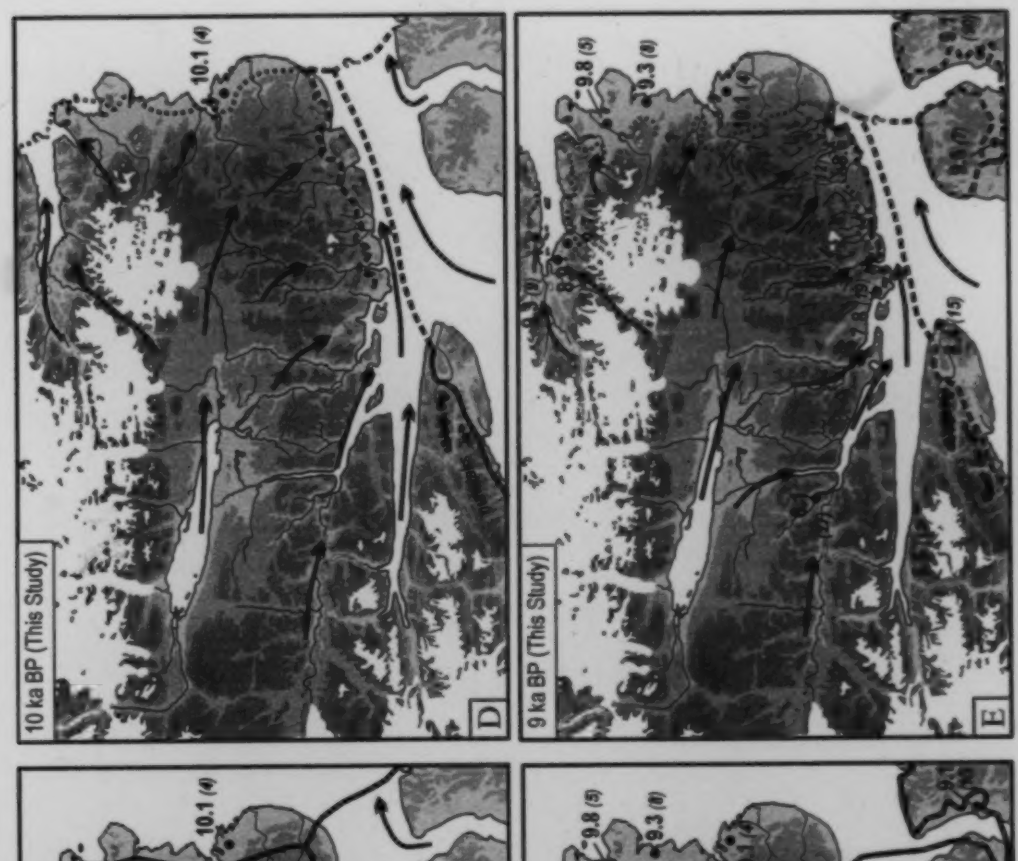

Last Glacial Maximum
Clements Markham Inlet
Piper Pass
Grant Ice Cap
?
Lincoln Sea
Alert
?
Lincoln Bay
Wrangel Bay
Beaufort Valley
South Basin
Robeson Channel
Hall Land
Cape Baird
Bay
Lady Franklin
Conybeare Fd
Dodge River
Agassiz Ice Cap
Archer Fd
Judge Daly Promontory
nunataks
A
N
0 km 25

Figure 5.10A-E *(over)* Paleogeographic maps illustrating ice configurations during the LGM and at 10 and 9 ka BP. Thick solid and dashed lines represent proposed ice margins and coalescent boundaries; lines with arrows indicate direction of ice flow; small barbed lines represent lateral meltwater channels (barb points upslope and downflow); shaded contour interval - 300 m. Key radiocarbon dates are indicated in thousands of years before present (ka BP) with site number from Table 5.1 shown in brackets. **5.10A** During the LGM, Agassiz ice crossed outer Hazen Plateau along a coalescent boundary with Greenland ice, the margin of which lay an unknown distance inland. Grant Land Mountain ice flowed northeast across Hazen Plateau, and northwards through Piper Pass. **5.10B-C** *(over)*. Paleogeographic maps for 10 and 9 ka BP proposed by England (1998, in press) illustrating a coalescent Ellesmere-Greenland boundary on Hazen Plateau, drawn along the western limit of Greenland erratics. Flow is convergent towards the coalescent boundary. **5.10D** Initial deglaciation of northeastern Ellesmere Island accompanies the offshore movement of the coalescent Ellesmere-Greenland boundary. Retreat of Grant Land Mountain ice on Hazen Plateau may have been initiated by this time, recorded by lateral meltwater channels that descend towards Robeson Channel. Retreat and flow of the trunk ice occupying Robeson Channel is record by lateral meltwater channels that descend northeastward along the outermost highlands. **5.10E** Retreat and thinning of trunk ice from the head of Robeson Channel, exposes a large central plain on Hall Land (Greenland). Separation between Agassiz ice (forming the western margin of the trunk glacier) and local plateau ice caps and Grant Land Mountain ice occupying the valleys, occurs along outermost Hazen Plateau - recorded by kames, kame terraces and deltas, originally interpreted to be of Greenland origin (Retelle 1986), but herein attributed to Ellesmere (Agassiz) ice reworking Greenland erratics. Continued retreat of Grant Land Mountain ice occupying Hazen Plateau is recorded by *en echelon* lateral meltwater channels descending towards the outer fiords and channels.


10 ka BP (This Study)
10.1 (4)
D
9 ka BP (This Study)
9.8 (5)
9.3 (8)
10.1 (4)
E
10 ka BP (England 1998)
10.1 (4)
B
9 ka BP (England, in press)
9.8 (5)
9.3 (8)
10.1
9.6 (7)
C

al. 1987) and from marine and geomorphic records on Greenland (Kelly & Bennike 1985; Funder 1989) suggests ice was thicker during pre-Sangamonian (Illinoian?) time. Whether increased ice thicknesses would result in a westward displacement of the coalescent Ellesmere-Greenland boundary is unknown. It is likely that dynamics along marine-based margins in the Lincoln Sea may be more significant.

5.6.2 Deglaciation of Robeson Channel and Northeast Hazen Plateau

Retreat of the trunk glacier occupying Robeson Channel is constrained by a date of 10 100±200 BP near Alert (Fig. 5.10B, Table 5.1). In the paleogeographic maps of England (1998, in press) the 9 ka BP coalescent Ellesmere-Greenland ice boundary along Hazen Plateau, remains essentially the same as that depicted at 10 ka BP (Fig. 5.10B, C). These reconstructions are questioned by three lines of evidence. The first is the dates of 9580±140 and 9070±150 BP (Table 5.1) from sites on Greenland across from Wrangel Bay (Fig. 5.10C). Ice flowing northwards out Robeson Channel would have a level profile, perpendicular across the channel, and ascend upflow towards the outlet of the Petermann Glacier in Hall Basin. Thus, it is not feasible to maintain Greenland ice on Hazen Plateau at elevations >600 m asl, while a shallow plain stretching across Hall Land (abutting the main flow of the Petermann Glacier), was deglaciated as early as 9.6 ka BP (Fig. 5.10C). The 9.6 ka BP age was measured on paired *P. arctica* shells in marine silts at 110 m (relative sea level <140 m; Table 5.1; England 1985). Other samples within this central plain (surface collections of paired *Hiatella arctica* in growth position) date between 9.0 and 9.2 ka BP (Weidick 1978; England 1985).

The second line of evidence concerns the nature and position of the coalescent boundary itself. In England's (1998) depiction of the 10 ka BP coalescent Ellesmere-Greenland boundary, ice flow (arrows) is indicated as being radial towards the suture of the two ice masses (Fig. 5.10B). While ice is shown to flow northwards out of Robeson Channel, there is no indication of a parallel flow along both sides of the suture. This would mean that the suture marks the lowest point along the trunk glacier, that the convergent Ellesmere and Greenland ice flow along this line indicates it was aggrading, and that at the same time, trunk ice was retreating from the mouth of Robeson Channel and the northeastern

Ellesmere Island mainland. Implicitly, the lowest point on a trunk glacier should correspond to the area of maximum ablation. Therefore, the 10 and 9 ka BP depictions of England (1998, in press; Fig. 5.10B, C) cannot be accurate, as this would mean ablation was greatest on land, rather than along calving margins in Robeson Channel/Lincoln Sea. When Ellesmere and Greenland ice were coalescent on top of the Hazen Plateau, the margin of the trunk ice must have lain offshore, and therefore, the region around Alert could not have been deglaciated.

The final line of evidence further questions the assertion that at 9 ka BP, Ellesmere-Greenland ice remained coalescent on top of Hazen Plateau (Fig. 5.10C). In Clements Markham Inlet, Bednarski (1986) reports several radiocarbon dates >9 ka BP, including a date of 9745±255 BP on *in situ* shells near the head of the inlet, that clearly document a deglacial, relatively restricted (onland) ice cover by 9 ka BP (Fig. 5.10E). When there was a coalescent Ellesmere-Greenland boundary on Hazen Plateau (England 1998, in press; Fig. 5.10A), Grant Land Mountain trunk ice would have overrun large sections of Hazen Plateau >900 m asl. This would equate to ice surfaces along the front of the Grant Land Mountains (the source of the trunk ice) and at the head of Piper Pass, >1000 m asl. Thus, Piper Pass (274 m asl) would have channelled a considerable ice flow northwards into the head of Clements Markham Inlet. Further, glaciers flowing into the head of Clements Markham Inlet are fed by the same icecap that supplied the trunk ice occupying Hazen Plateau, and therefore are themselves likely to have expanded. Thus, it seems unlikely that ice margins could be so restricted in Clements Markham Inlet, when Ellesmere and Greenland ice were still coalescent on outer Hazen Plateau (Fig. 5.10B, C).

Alternative models of ice cover are thus proposed. During the LGM, a trunk glacier flowed northeastwards beyond the head of Robeson Channel (England, in press). Ellesmere and Greenland ice were coalescent along a suture that lay some unknown distance (≤15 km) inland of the outer Hazen Plateau, along which a parallel flow of Ellesmere (Agassiz) and Greenland ice occurred (Fig. 5.10A). During deglaciation the suture moved coastward prior to the exposure of local summits as the uppermost lateral meltwater channels descend towards Robeson Channel, rather than descending inland. Movement of the suture offshore would have occurred prior to establishment of the 10.1 ka BP date at Alert. Continued thinning of the trunk ice occupying Robeson Channel, led to a separation between it and local ice atop Hazen Plateau. Previously, ice contact features (kames, kame terraces and deltas) mapped along the outer Ellesmere Island coast were considered to mark the separation of Greenland ice occupying Robeson Channel and local (Ellesmere) plateau ice caps (Retelle 1986). These included kames at 415 m asl on the upland slopes west of South Basin deposited by the Greenland ice, and a large delta at 397 m that prograded from local plateau ice into a proglacial lake impounded between the two ice masses (Fig. 5.10E; Retelle 1986). Below this, Retelle (1986) documented the continued thinning and lobate flow of the Greenland ice along the coast by successive lateral and end moraines, kames and deltas that prograded inland, and lateral meltwater channels that ringed the outermost highlands, descending northeast. Greenland ice is invoked as the source of these landforms because they all contain "abundant Greenland-type erratics" (Retelle 1986). However, under the reconstruction proposed here, the trunk ice flowing immediately along the outer Ellesmere Island coast would have been Ellesmere (Agassiz) ice, and therefore, requires that the Greenland erratics be redeposited by Ellesmere ice overriding Greenland till. The age of these ice-marginal deposits is unknown, but may be correlative with the initial retreat of trunk ice from the head of Robeson Channel (~60 km northeast). The date of 10.1 ka BP from Alert which records deglaciation of the northeastern Ellesmere Island mainland, may thus provide a limiting age for this. On outer Hazen Plateau, plateau ice caps may have been widespread, while abundant lateral meltwater channels record the thinning and retreat of Grant Land Mountain ice northwest up the regional valleys (Fig. 5.10E).

The lowest landforms recording continued occupation of Robeson Channel by trunk ice are an undated delta complex at 220 m asl in the Beaufort Valley that progrades inland, and several small kames ~191 m asl along the eastern slopes of Beaufort Valley and South Basin (Retelle 1985, 1986). The absence of subsequent deposits below this, likely indicates a rapid breakup of the trunk ice. Although undated, the age of the delta probably falls between the 10.1 ka BP date at Alert, and the >9 ka BP dates on Hall Land, if geometry of the trunk ice and deglaciation of the plain across Hall Land are taken into consideration. The absence of dates >9 ka BP in Lincoln Bay is interpreted to reflect continued blocking of the sea by trunk ice in Robeson Channel (England, in press). Whether this occurred from grounded trunk ice, or an iceshelf is unknown, although the later may better explain the asymmetry of ice across the channel, and accommodate an ice margin well inside central Hall Land by ~9.6 ka BP. Retelle (1985) also notes that in Lincoln Bay, the 8.6 ka BP date collected from 91 m asl is a minimum estimate of a poorly defined washing limit at ~100 m asl, in a steeply cliffed region where marine limit is often indeterminate. Thus, the possibility exists that Lincoln Bay was deglaciated by 9 ka BP.

5.6.3 Final Deglaciation

Thinning and retreat of the Robeson Channel trunk ice led to the deglaciation of much of Hazen Plateau, and the increasing confinement of Grant Land Mountain trunk ice within the regional valleys and fiords. Plateau ice caps, however, continued to persist. Indeed, cross-cutting meltwater channels north of Hart Lake (Fig. 5.5) indicate that plateau ice caps expanded locally during the early Holocene. Retelle (1986) also cites evidence of early Holocene growth of plateau ice caps along Robeson Channel, where they contacted the sea, forming marine-limit deltas. The growth of plateau ice caps at a time when marine-based ice margins were retreating indicates major differences in mass balance and stability. Accumulation records from the Greenland Ice Sheet, point to a 2-5 fold increase in accumulation immediately following the Younger Dryas, resulting principally from changes in atmospheric circulation (as opposed to temperature; Alley *et al.* 1993; Kapsner *et al.* 1995). Growth of ice in the High Arctic may thus have been favoured. The retreat of marine-based ice at this time, may also reflect a dominant eustatic control. Opening of Lady Franklin Bay and separation of the Ellesmere-Greenland suture is recorded by dates of 8.5 ka BP midway up Archer Fiord, and 8.4 ka BP on the Cape Baird Terrace (Fig. 5.10E; Table 5.1; England, in press). By 8 ka BP, retreating Grant Land Mountain trunk ice was grounded at the mouths of many of the fiords and valleys throughout the field area, and continued to occupy Lake Hazen Basin and the inner Hazen Plateau (Fig. 5.9). Once ice moved inland of Hazen Plateau and west of the central Grant Ice Cap, it began to impound the regional drainage, forming extensive proglacial lakes. This would have likely accentuated retreat rates, as ice margins began to calve, or in the example near the head of The Bellows Valley (Fig. 5.2B), were quickly destabilized and eroded by spillways. At 7 ka BP, ice margins appear to have stabilized within Lake Hazen Basin (Fig. 5.9), retreating only slightly between then and 6 ka BP, reflecting in part, both a completely land-based margin, and free drainage away from the ice. Plateau ice caps persisted well into the mid-Holocene, meltwater from which built deltas in several local basins. Elsewhere within north-central Ellesmere Island, England (1986) also sites evidence of the persistence of plateau ice caps through the mid-Holocene, suggesting that regional ELA levels may have remained considerably depressed. The physical presence of glaciers occupying Lake Hazen Basin would also have acted as positive feedback on ice stability, as it would reduce some of the intermontane characteristics of the Lake Hazen region, responsible for the summer thermal oasis conditions seen today. Almost all of Lake Hazen Basin appears to have cleared of ice by 5 ka BP, possibly in as little as 300 years as suggested by the 5.3 ka BP date in Connell Lake (Fig. 5.9; Table 5.1). Following this retreat, paleoenvironmental records from lake cores throughout the field area (Chaper 2 of this thesis) signal a rapid increase in diatom productivity supporting the establishment of conditions similar to those of today.

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THE DEGLACIAL AND HOLOCENE STABLE ISOTOPE STRATIGRAPHY OF THE FORAMINIFERA *Lobatula lobatulus*, FROM AN ISOLATION BASIN, HOVED ISLAND, CANADIAN HIGH ARCTIC

6.1 INTRODUCTION

Stable isotope records of marine organic carbonates, such as the tests of foraminifera and shells of bivalve molluscs, have become increasingly useful in understanding the nature and style of deglaciation (Epstein & Mayeda 1953; Hillaire-Marcel 1981; Mix & Ruddiman 1985; Andrews *et al.* 1991, 1993). Temperature variations of Arctic waters during the last glaciation have generally been considered small (1 - 4 °C), accounting for at most, a 1‰ change in the oxygen isotope record (van Donk & Mathieu 1969; Aksu & Vilks 1988; Israelson *et al.* 1994). Therefore, variations in the stable oxygen isotope records of Arctic marine carbonates dominantly reflect changes in salinity caused by mixing with isotopically light glacial meltwater (Craig & Gordon 1965; Vilks & Deonarine 1988; Spielhagen & Erlenkeuser 1994). Changes in the stable carbon isotope record are linked to changes in sea ice cover, planktonic productivity, and the incorporation of terrestrially-derived organic carbon and dissolved inorganic carbon (DIC) (Kroopnick *et al.* 1977; de Vernal *et al.* 1992; Mackensen *et al.* 1993). The quality of the deglacial isotopic record is variable, and often relates to the local hydrography and paleo-depth at which the organisms existed. In the open ocean and deeper straits and bays, benthic organisms may show little isotopic dilution, while their planktonic counterparts that occupied an upper, dilute surface layer, exhibit considerably greater isotopic variations (Erlenkeuser 1985; Scott *et al.* 1989; Andrews *et al.* 1994). In areas of restricted circulation, or shallow shelf regions, the isotopic shifts of benthic organisms may be considerably amplified (Olausson 1982).

Many fiords and inter-island channels within the Canadian Arctic Archipelago are characterized by restricted circulation and/or shallow water depths (Fissel *et al.* 1984). Planktonic foraminifera are largely absent from the marine sedimentary records from this region. However, benthic species are relatively abundant (Schröder-Adams *et al.* 1990; Scott & Vilks 1991). This study investigates the deglacial and Holocene stable isotope records of the epibenthic foraminifera *Lobatula lobatulus* (Walker & Jacob) (= *Cibicides lobatulus)*, and the infaunal mollusc *Portlandia arctica* (Gray) within a sediment core from an isolation basin on Hoved Island, Canadian High Arctic (Fig. 6.1).

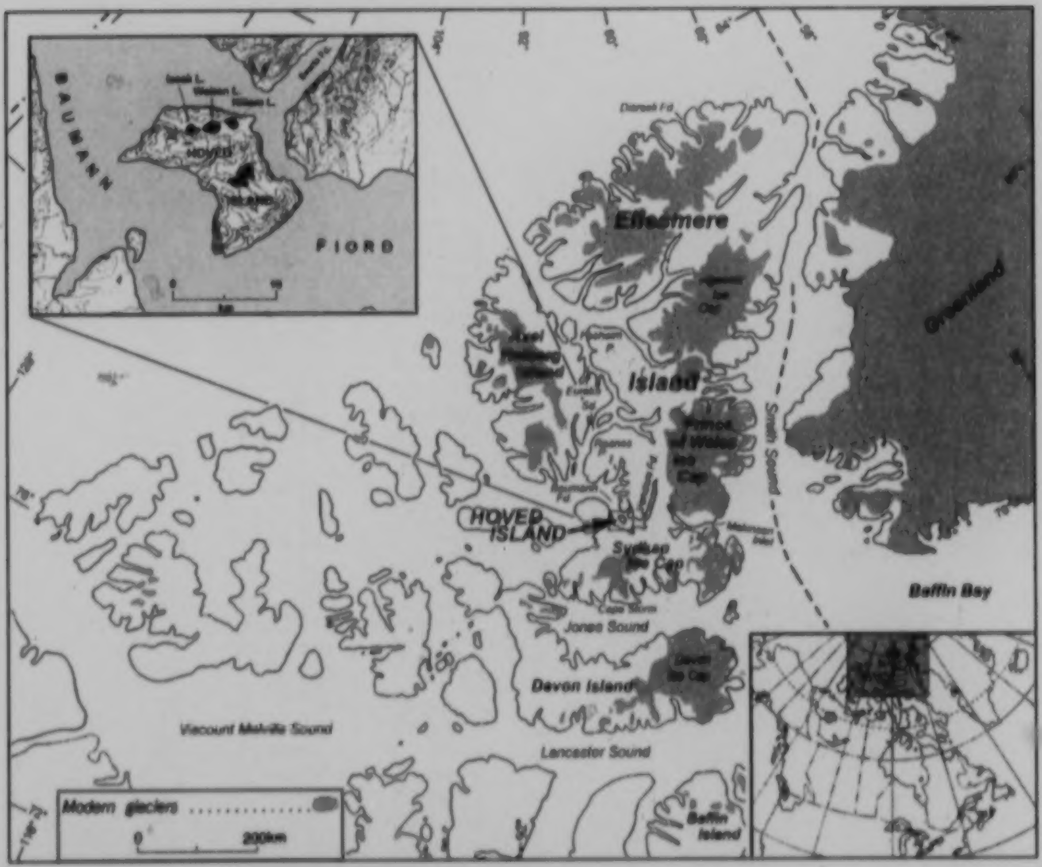
BAUMANN
HOVED
FIORD
Ellesmere
Island
Prince of Wales Ice Cap
Smith Sound
Greenland
HOVED ISLAND
Sydkap Ice Cap
Jones Sound
Devon Island
Devon Ice Cap
Lancaster Sound
Baffin Bay
Baffin Island
Viscount Melville Sound
Modern glaciers
0
200km

Figure 6.1 Study area on southwest Ellesmere Island, within the Canadian Arctic Archipelago (inset, lower right). Upper left inset shows detail of Hoved Island, its lakes (darkly shaded), and adjacent areas; contour interval is 200 ft.

6.1.1 Study Area

Hoved Island is situated in central Baumann Fiord, southwest Ellesmere Island (77° 32' N, 85° 14' W, Fig. 6.1). It is crossed by three steep-sided ridges and has a maximum elevation of 302 m above sea level (asl). There are five lakes on the island, four of which are below Holocene marine limit (98 m asl). Three lakes are linked at the north end of the island (*Izaak*, *Walton* and *Killam* Lakes, unofficial names; Fig. 6.1), the water surfaces of which are 5.5, 1.5 and 0 m asl, respectively. Izaak and Walton lakes are considered isolation basins; that is, they were initially transgressed and subsequently isolated from the sea as a result of glacioisostatic adjustments following deglaciation. Killam Lake is still connected to the sea by a 3 m deep outlet channel, but will eventually become isolated due to ongoing postglacial emergence. The island is covered by a veneer of weathered till containing granite erratics. Mollusc shells also occur as erratics on surfaces up to 200 m asl.

6.1.2 Physiography

Ellesmere Island in the vicinity of Hoved Island is characterised by ridges and valleys >600 m in relief. These are developed in Upper Proterozoic and Paleozoic carbonate and clastic rocks of the Franklinian mobile belt which strikes NE-SW (Trettin 1989). The same structural trends are seen in the bathymetry of Baumann Fiord, where water depths exceed 300 m, and in adjoining Vendom Fiord, where depths are as great as 450 m. Outcrops of Devonian coal-bearing strata occur extensively southwards of central and inner Baumann Fiord, and also along the west shore of Vendom Fiord (Ricketts & Embry 1984). Tertiary coal-bearing strata of the Eureka Sound Group outcrop at the mouth of Vendom Fiord and along the south-central and inner shore of Baumann Fiord (Ricketts & Embry 1984). Highly metamorphosed sedimentary rocks and igneous rocks of the Canadian Shield (diagnostic erratics in this region), outcrop ~100 km to the east, in the mountains of eastern Ellesmere Island. These rocks are mostly covered by the Prince of Wales and Agassiz Ice caps.

6.1.3 Glacial History

Hodgson (1985) studied the surficial geology of central and western Ellesmere Island where he noted the presence of a prominent "drift belt" extending 10-60 km beyond present-day ice caps. This drift belt is generally confined to the fiord heads and represents the margins of expanded and coalesced ice caps dating between 9 and 7 ka BP. Within the central and inner regions of Baumann Fiord, the drift belt was formed by ice advancing from three principal areas: 1) Braskeruds Plain and Svendsen Peninsula (between Raanes Peninsula and the Prince of Wales Ice Cap; Fig. 6.1), 2) the divide between Vendom Fd. and Makinson Inlet, and 3) the southwestward extension of this divide, presently supporting the Sydkap Ice Cap (Fig. 6.1; Hodgson 1985).

Whether or not the drift belt marked the limit of glaciers during the LGM was uncertain (Hodgson 1985). Previous work on and around Fosheim Peninsula (Fig. 6.1) would argue that it does, suggesting that the LGM was characterised by a discontinuous ice cover, constrained by severe aridity and calving (Bell 1996; Bednarski 1995; England 1990, 1992; Hodgson 1985, Model B; Lemmen *et al.* 1994). However, recent investigations by England (1989, in press), Bednarski (in press), Dyke (in press) and O'Cofaigh (1998), question the evidence on which past models of restricted ice margins were based, and present evidence indicating that glaciers inundated the high Arctic islands and marine channels during the LGM. Based on these investigations it appears likely that Hoved Island was covered by ice during the LGM.

6.1.4 Oceanography

The oxygen isotope composition of marine waters within the Canadian Arctic Archipelago is strongly controlled by local bathymetry, the influx of water from the Arctic Ocean, sea ice melt, and the input of isotopically depleted meteoric waters (Tan & Strain 1980a; Bédard *et al.* 1981; Strain & Tan 1993).

The Archipelago itself is composed of numerous islands, surrounded by shallow, narrow waterways and sills, separating deeper basins. Water from the Arctic Ocean, driven by the Beaufort Sea Gyre and the Transpolar Drift, is channelled through the Archipelago, emptying into Baffin Bay (Bailey 1957; Ostenso 1966). Three distinct water masses have been recognized in the Archipelago, the uppermost of which is termed the Arctic or surface water mass (Fissel *et al.* 1984). Beneath this is a cold isohaline water mass, extending to depths of 200 m in most places, and as much as 300 m in the deeper straits such as Lancaster

Sound (Hunt & Corliss 1993). An Atlantic water mass underlies this within the Arctic Ocean, but is excluded from the Archipelago by numerous shallow sills (Collin & Dunbar 1964). Temperature, salinity and isotopic composition within the upper few tens of metres of the surface water mass are highly variable. Along Baffin Island this surface water mass has temperatures of 0 to 8°C, salinity from 20 to 30‰, and develops to depths of 5 to 60 m in the fiords, and 50 to 100 m across the continental shelf (Ellis 1960; Gilbert 1978; Trites 1985). Within Jones Sound, Tan & Strain (1980b) have recorded $\delta^{18}O_{SMOW}$ values ranging from < -2.37 ‰ at the surface to ~ -2.14 ‰ at 10 m depth and ~ -1.27 ‰ at 140 m depth. Using the reported temperatures and $\delta^{18}O_{SMOW}$ values (corrected for an acid fractionation factor of -0.22 ‰; Tan & Strain 1980b), and employing the paleotemperature equation published by Shackleton (1974);

$$T = 16.9 - 4.38(\delta_c - \delta_w) + 0.10(\delta_c - \delta_w)^2$$

the $\delta^{18}O_{PDB}$ values of inorganic carbonate (δ_c) forming in equilibrium with these waters (δ_w) can be estimated as: 1.75, 1.98 and 3.06 ‰, respectively.

The $\delta^{18}O_{SMOW}$ composition of meteoric waters, and hence glaciers, within the Arctic Archipelago is isotopically very light (-20 to -35 ‰; Koerner 1979). Consequently, meltwater inputs can result in large isotopic shifts within the marine records (Andrews *et al.* 1993). For example, meltwater from Alfred Wegener Glacier, west Greenland, has a $\delta^{18}O_{SMOW}$ composition between -20.5 and -21.7 ‰ (Bédard *et al.* 1981). This mixes rapidly with the surface waters producing values of -16 ‰ at the outlet of the glacier increasing to -6.3 ‰ at the mouth of the fiord. Similarly, $\delta^{18}O_{SMOW}$ values increase rapidly with depth, averaging -1.3 ‰ at 20 m and -0.8 ‰ at 50 m. Assuming temperatures of 0°C, and using the same calculations as above, inorganic carbonate precipitated in equilibrium with these waters would have $\delta^{18}O_{PDB}$ values of approximately: -11.95, -2.24, 2.74 and 3.24 ‰, respectively.

During deglaciation, the circulation through the Canadian Arctic Archipelago would have differed from that of today. Many channels in the southern islands were infilled by the Laurentide Ice Sheet, which discharged large volumes of meltwater northwestward, reversing the present-day surface flow through the Archipelago (Fig. 6.2; Dyke & Prest 1987; Andrews *et al.* 1993). This would have strongly altered the salinity and isotopic composition of the upper dilute surface layer, which would have extended considerably deeper than that of today. A tentative reconstruction of circulation patterns in the Archipelago for 8.5 ka BP and the present is provided by Dyke & Morris (1990; Fig. 6.2A). The establishment of the present surface water circulation (Fig. 6.2B) is considered to have occurred shortly after 8.5 ka BP, based on two lines of evidence (cf. Dyke *et al.* 1996, 1997). Abundant remains of bowhead whales (*Balaena mysticetus*), dated between 10.5 and 8.5 ka BP, are found throughout much of the Archipelago (Dyke & Morris 1990; Dyke *et al.* 1996), whereas, initial driftwood entry occurs ~8.6 ka BP. Therefore, driftwood must have been excluded before 8.6 ka BP by something other than local sea-ice conditions or glacial cover. The switch from an Archipelago circulation (NW) driven by glacial meltwater to the present southeastward flow of Arctic Ocean surface water is recorded by the import of Arctic Ocean sea ice, which carried driftwood, and simultaneously, restricted the bowhead whale range (Dyke *et al.* 1996). Driftwood within Eureka Sound, however, is rare until approximately 6 ka BP (Stewart & England 1983; Dyke *et al.* 1997). It is uncertain whether this relates to a continued dominance by glacial meltwater outflow and an even later establishment of modern surface currents, or whether it reflects the Holocene sea ice (fast ice) history.

Ice shelves, including both sea-ice and glacial ice shelves (Jeffries 1992), may also strongly control isotopic composition of surface water. Within Disraeli Fiord (Fig. 6.1), Keys *et al.* (1969) reported that behind the 44 m thick ice shelf, the inner fiord waters were fresh down to a depth of 44 m. Between 44 and 45 m, the salinity increased from 5 to 32 ‰, while the temperature decreased from -0.6 to -1.6°C. Based on the calculations of Jeffries *et al.* (1989), this would correspond to a $\delta^{18}O_{SMOW}$ shift from approximately -25.1 to -2.1 ‰.

Today, the only ice shelves in the Canadian Arctic Archipelago occur on northern Ellesmere Island. Sea-ice ice shelves are, however, ephemeral phenomena (Stewart & England 1983; Evans 1989) and were clearly more widespread earlier this century (Koenig *et al.* 1952). Evidence of former glacial ice shelves has been documented along eastern Ellesmere Island (England *et al.* 1978), and along the NW margin of the Laurentide Ice Sheet in Viscount Melville Sound (Hodgson & Vincent 1984; Dyke 1987; Hodgson 1994).

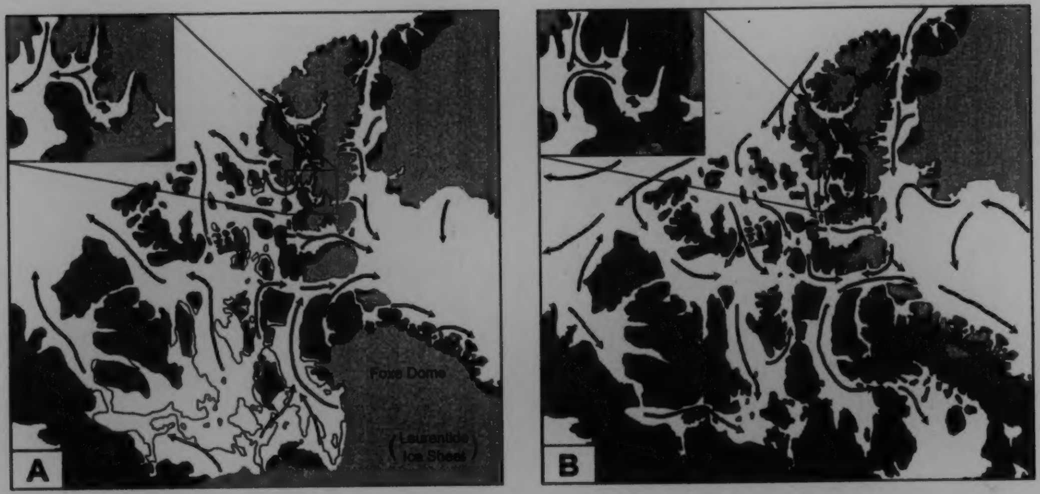

Figure 6.2 Circulation patterns through the Arctic Archipelago. **6A** Proposed ocean surface currents ca. 8500 BP (*after*, Dyke & Morris 1990, their Fig. 6, redrawn with permission of the Geological Survey of Canada). Ice margins approximate those of Hodgson 1985, and Dyke & Prest 1987. Outlines of modern shorelines of the Arctic islands illustrates the relative extent of land submerged below sea level at this time. Inset shows detail of the Baumann Fiord region. **6B** Modern ocean surface currents of the Canadian Arctic Archipelago (*after*, Ostenso 1966).

6.2 METHODS

6.2.1 Sediment Cores

The three linked lakes (Izaak, Walton and Killam, Fig. 6.3) were cored in the spring of 1994 using a vibracorer system (Smith 1992). This coring system results in very little core-side disruption (< 5 mm) and downwarping of the sediment, but does cause varying amounts of sediment compaction. Single cores were taken from the deepest basin of each lake, utilizing the 2.1 m thick ice pan as a coring platform. Core lengths from Izaak, Walton and Killam lakes were 5.77, 3.00 and 6.33 m, respectively (Fig. 6.4); sediment compaction was 0.90, 0.45 and 0.75 m, respectively. Cores were sectioned into 2.25 m lengths in the field, capped, frozen and shipped south. They were then split lengthwise, while frozen, with a diamond saw at Core Laboratories (Calgary). The sediments of all three cores were predominantly a silty, marine, bioturbated mud, containing whole molluscs (mostly *Yoldiella* spp. and *P. arctica*), abundant foraminifera and ostracodes, and occasional ice-rafted pebbles. The cores from Izaak and Walton lakes bottomed out in a coarse diamict (Figs. 6.4 and 6.5).

6.2.2 Radiocarbon Dating

Six samples were submitted for AMS radiocarbon dating to IsoTrace Laboratories, University of Toronto (Table 6.1). Of these, two were lost in the HCl leach designed to remove surface contamination (TO- 4760 and TO-4761; Walton Lake core, 250-255 and 260-265 cm depth respectively). Most samples included several paired molluscs, as individuals were generally too small to provide sufficient carbon for analysis. The oldest date shows continuous sedimentation in the Killam Lake basin, back to at least 9.3 ka BP (Table 6.1).

6.2.3 Stable Isotope Analysis

Continuous 5 cm increments (representing ~150 to 350 years of sedimentation) were cut from half sections of the Walton Lake core. The outer 5 mm of material was removed to avoid contamination from material smeared along the walls of the coring tube. Sample volume was determined by displacement. Foraminifera and molluscs were separated by washing through a 63 μm sieve. Individual *L. lobatulus* tests and *P. arctica* shells were

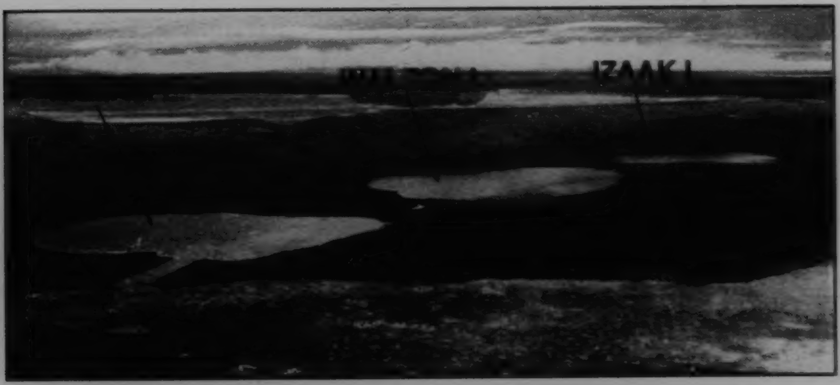

Figure 6.3 Oblique photo showing the study lakes, situated between two ridges on the north end of Hoved Island. View is to the south-southwest.

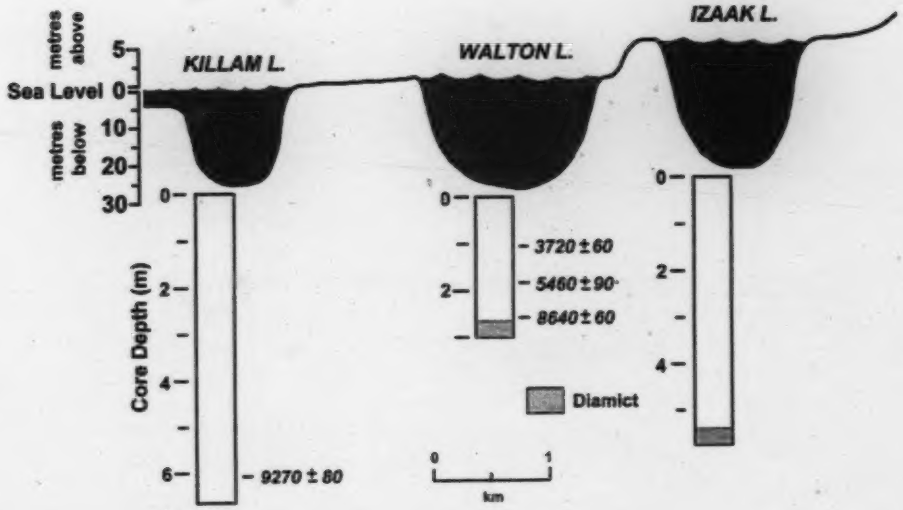
IZAAK L.
metres above
5
KILLAM L.
WALTON L.
Sea Level 0
metres below
10
20
30
0
2
4
6
Core Depth (m)
9270 ± 80
0
2
3720 ± 60
5460 ± 90
8640 ± 60
Diamict
0
1
km
0
2
4

Figure 6.4 Sediment core logs and spatial distribution of the three study lakes. Radiocarbon dates on the figure are listed in Table 6.1

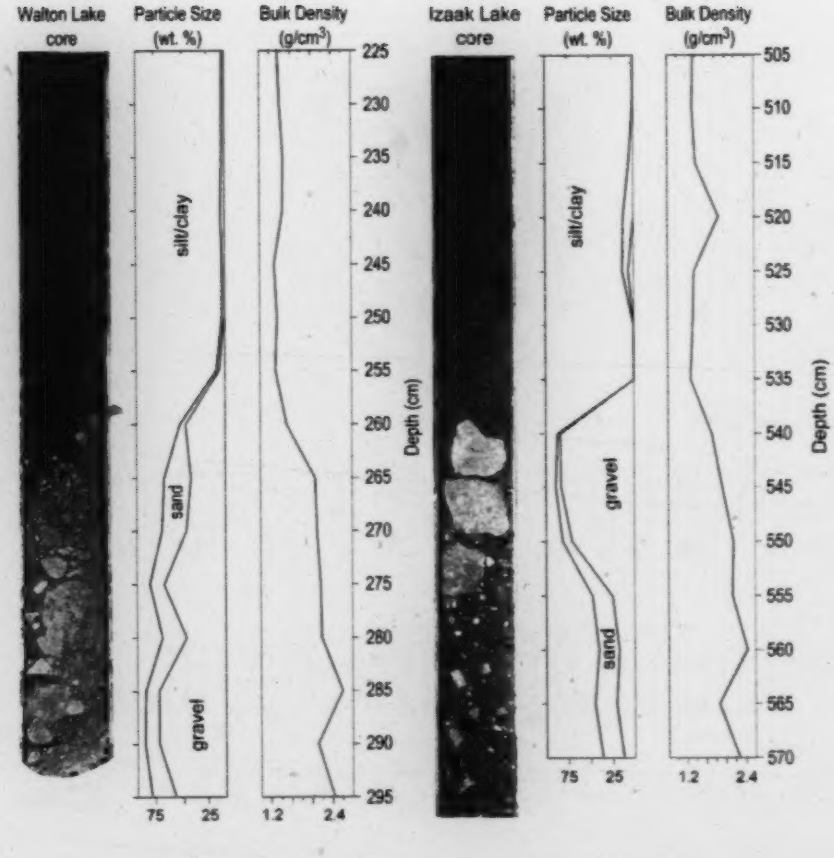

Figure 6.5 Lowermost core sections from Walton and Izaak lake, showing the coarse diamicts, and their particle size and bulk density (dry) characteristics. Bulk density was determined on 2 cc. samples taken with a cut syringe, and dried at 105°C. Bulk density within the coarse diamict is of the matrix material, and was packed into the syringe by hand.

Table 6.1 Radiocarbon date list.

^{14}C age (BP)	Laboratory No.	Material	Lake	Core Depth (cm)	Enclosing sediment
9270±80[a]	TO-4757	1 *Portlandia arctica* (pd, gp)[b]	Killam	608	marine silt
8640±60	TO-4929	2 *Portlandia arctica* (pd)	Walton	260-265	marine silt with dropstones
5640±90	TO-4759	10 *Portlandia arctica* (pd); 1 *P. arctica* (s)	Walton	190-200	marine silt
3720±60	TO-4758	6 *Portlandia arctica* (pd, gp)	Walton	105-110	marine silt

[a] Dates are reported with an ocean reservoir correction of 410 years, following normalization of $\delta^{13}C$ to a base of -25 ‰. Errors represent ± 1 standard deviation (Beukens pers. comm. 1996).

[b] Abbreviations for molluscs are: pd - paired; gp - growth position; s - single valve

hand-picked from the sample. Not all 5 cm sections had sufficient number of tests for analysis.

Samples were cleaned in a sodium hypochlorite solution by ultrasonification for 5 minutes, and then left to soak for 24 hrs to remove organic material and the periostracum (Durazzi 1977; Stevens & Vella 1981; Israelson *et al.* 1994). They were then repeatedly washed in distilled water, left to soak overnight, and finally rinsed with acetone and dried at 60°C. Prior to analysis, the foraminifera tests were gently broken, and the internal chambers were inspected to ensure that no errant mineral grains were present. Tests and shells showing any visible staining, etching or concretions were discarded.

Generally five *L. lobatulus* tests were used in each analysis, although as few as three and as many as seven were used in some samples. Replicate samples from the same 5 cm core segment were run to assess intra-sample variability (Table 6.2). Studies have demonstrated isotopic variations within different size fractions of the same foraminifera species (cf. Berger *et al.* 1978). However, because of low numbers of tests throughout much of the Walton Lake core (Fig. 6.6) it was not possible to discriminate samples on this basis. Single, whole valves of *P. arctica* were analyzed.

Samples were reacted with 100% phosphoric acid, and the evolved CO_2 gas was cryogenically captured and analyzed on a Finnegan MAT 252 mass spectrometer. Isotopic composition is recorded in standard delta (δ) notation in per mil (‰) relative to the Peedee belemnite (PDB) international standard (Epstein *et al.* 1953). Isotopic compositions reported are the average of 10 individual measurements by the mass spectrometer. Eleven samples, for which the standard deviation of the 10 individual measures was greater than 0.1 ‰, were discarded (higher measurement standard deviations are considered the result of low gas pressures relating to small sample size and possible fractionation of the gas within the capillary tubes).

Table 6.2 Stable isotope concentrations (‰ PDB) for replicate samples of *Lobatula lobatulus*, Walton Lake core.

Sample Depth (cm)	# of foram tests per sample	$\delta^{13}C$	$\delta^{18}O$
90 - 95	5	-4.34	-6.34
90 - 95	5	-0.88	-3.39
110-115	5	-12.30	-9.02
110-115	5	-2.10	-3.95
135-140	5	-4.51	-7.93
135-140	5	-3.92	-7.64
140-145	5	-17.90	-20.90
140-145	6	-15.20	-18.80
140-145	5	-0.30	-3.24
140-145	5	0.46	-0.60
155-160	5	-3.49	-5.58
155-160	5	-0.37	-1.48
160-165	5	-6.98	-9.58
160-165	4	-2.40	-4.29
160-165	4	-3.02	-4.78
165-170	5	-5.28	-8.41
165-170	5	0.35	-0.19
165-170	7	0.00	-0.66
175-180	5	-4.91	-5.98
175-180	5	-3.02	-4.35
175-180	6	-2.40	-3.37
175-180	5	-0.92	-1.24
175-180	5	-0.23	-0.85
205-210	5	-1.60	-2.36
205-210	4	-1.00	-2.15
210-215	5	-15.33	-12.95
210-215	5	-6.85	-10.99
255-260	5	-0.22	1.45
255-260	5	-0.31	0.98
260-265	5	-8.70	-10.24
260-265	4	-0.27	0.81
260-265	5	-0.28	1.04
260-265	5	-1.81	-1.37
265-270	5	-3.74	-2.37
265-270	5	-6.68	-6.12

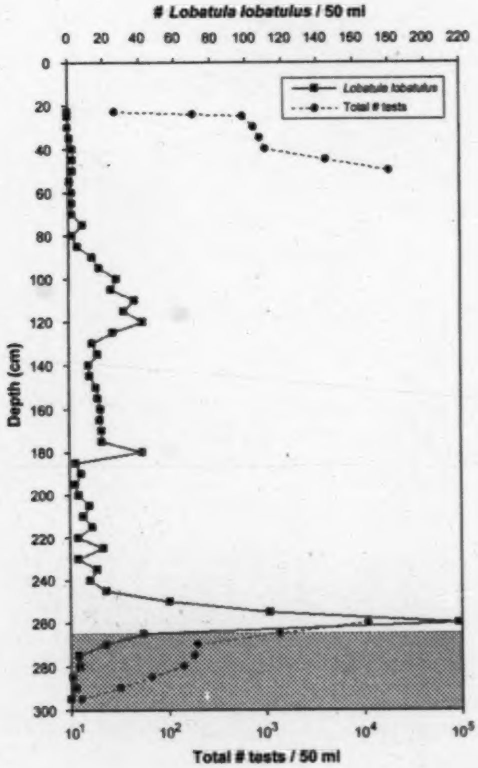

Figure 6.6 Number of *Lobatula lobatulus* tests per 50 ml of sediment, contrasted with the total number of foraminifera tests of all species per 50 ml, Walton Lake core. Note, the total number of foraminifera tests are plotted on a logarithmic scale. Shaded region represents the coarse diamict.

6.3 RESULTS AND DISCUSSION

6.3.1 Core Sedimentology

The basal diamict in the Walton Lake core extends from 300-265 cm where there is a gradual transition to massive marine silt with a single gravel interbed and one dropstone between 263-258 cm (Figs. 6.4 and 6.5). Within the Izaak Lake core, the diamict extends from 575-540 cm where there is an abrupt transition to massive marine silt, with a small accumulation of sand and gravel between 525-515 cm (Figs. 6.4 and 6.5).

There are two alternative interpretations of the genesis of the basal diamicts: 1) they are tills, deposited by grounded ice infilling Baumann Fiord, or 2) they were formed by debris rain-out from a glacial ice shelf, formed from a previously grounded glacier within Baumann Fiord. Both the isotopic records from Walton Lake, and the biological subfossil remains within the diamicts support the latter interpretation. Similar diamicts have been attributed to deposition by glacial ice shelves in both the Arctic and Antarctic (Anderson *et al.* 1991; Hodgson 1994; Powell 1994). With respect to the subfossil record, the fact that fragile yet unbroken foraminifera, ostracodes and paired molluscs are found within the diamicts does not preclude their redeposition by ice. However, the exponential increase in foraminifera (Fig. 6.6) and the increase in species diversity (3 - 295-300 cm; 39 - 265-270 cm) is typical of the colonizing pattern shown for ice shelves and associated marine deposits in Antarctica (Kellogg *et al.* 1979; Kellogg & Kellogg 1988; Domack *et al.* 1995). Foraminifera densities (all species) are initially low, 13 tests per 50 ml of sample (295-300 cm), but increase rapidly to 1378 per 50 ml (265-270 cm) and to 11,100 per 50 ml in the sediments immediately overlying the diamict (260-265 cm)(Fig. 6.6; note, counts include both adult and juvenile tests). Part of this exponential increase in densities likely relates to differences in sedimentation rate and the very coarse nature of the diamict (cobbles up to 6 cm in diameter). The absence of a basal diamict within the Killam Lake core suggests that it occurs below the depth of recovered sediment (therefore predating 9270±80 BP; Fig. 6.4).

Within the Walton Lake core, the isolation contact, marking the emergence of the lake basin out of the sea, occurs between 24 and 25 cm. This is identified by the dramatic decline in foraminifera numbers (Fig 6.6), the presence of freshwater diatoms above this point, and the stark change in colour (5YR 3/1 (m) to 7.5YR 4/2 (m)) relating to the change in sediment source of an open marine basin, to a terrestrially confined one. The age of the isolation contact is uncontrolled as insufficient material was found for dating. However, using the emergence curve of Blake (1975), from Cape Storm (~150 km to the south, Fig. 6.1), a crude estimate of 900 BP can be made for the 1.5 m asl outlet of Walton Lake.

6.3.2 Stable Isotopes

Stable isotope results, including the results of replicate analyses (Table 6.2), are plotted in Figures 6.7A, B and C.

6.3.2.1. *Vital Effects and Sample Variability*

The isotopic records derived from the epibenthic foraminifer *L. lobatulus* clearly differ from those of the infaunal mollusc *P. arctica* (Fig 6.7C). Isotopic differences between species and organisms from the same habitat are well known and may reflect species-dependant but otherwise constant "vital" effects (Duplessy *et al.* 1970; Hillaire-Marcel 1981; Dunbar & Wefer 1984; Taviani & Aharon 1989). Isotopic differences between species living infaunally (within the sediment) and those living epifaunally (on top of the sediment) may also occur as a result of microhabitat effects (McCorkle *et al.* 1990; Chandler *et al.* 1996). Hunt & Corliss (1993), studied the distribution of living benthic foraminifera within the Canadian Arctic Archipelago, and showed that many are most abundant below the sediment-water interface, down to 17 cm sediment depth, while more than half occur below 4.3 cm depth. They do not report on the habitat specificity of *L. lobatulus*. However, as noted by Corliss (1985), the plano-convex shape of the *L. lobatulus* test is specific to those species travelling on or near the surface, and is not suited to an infaunal existence. Furthermore, both Vilks & Deonarine (1988) and Poole *et al*. (1994) interpreted isotopic results from *L. lobatulus* tests as showing carbonate precipitation under near equilibrium conditions (Poole *et al.* 1994, report on an unpublished study showing a $\delta^{18}O$ disequilibrium value for *L. lobatulus* of -0.58 ‰). Differences between these two organisms may also relate to the timing of carbonate formation. Israelson *et al.* (1994) suggest that maximum carbonate shell formation may occur from June to August when food supply and light intensity are greatest. The isotopic composition of the shells would thus represent the temperature and $\delta^{18}O$

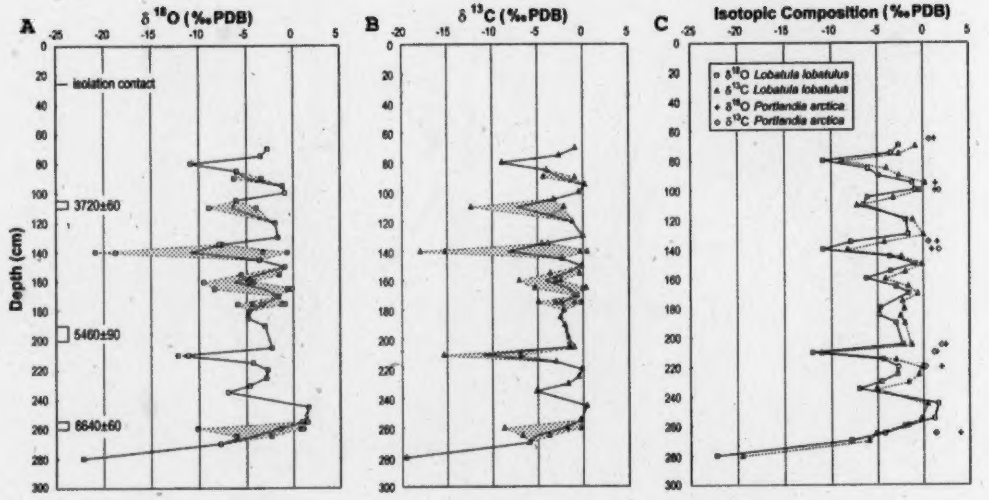

Figure 6.7 Stable isotope results from Walton Lake core. **6.7A** Oxygen isotope stratigraphy of *Lobatula lobatulus*, inclusive of the replicate samples, Walton Lake core. **6.7B** Carbon isotope stratigraphy of *L. lobatulus*, inclusive of replicate samples, Walton Lake core. **6.7C** Oxygen and carbon isotope stratigraphy of *L. lobatulus*, Walton Lake core, using average values from replicate analyses. Also shown are the oxygen and carbon isotope compositions of the *Portlandia arctica* shells.

conditions of the summer season. However, the fact that foraminifera share similar life strategies, and that the summer periods would relate to the most intense dilution by isotopically light glacial meltwater, argue that other factors must be responsible for the measured variations.

Studies of deep-sea sediments have also shown that the pore-water dissolved inorganic carbon (DIC) within the upper 1-2 cm, can be as much as 2-3 ‰ less than that of the bottom water (McCorkle *et al.* 1985; Sayles & Curry, 1988). Although *P. arctica* lives infaunally, it is a suspension feeder (Aitken 1990), and thus is less likely than *L. lobatulus* to be influenced by ^{13}C-depleted pore-water DIC.

There is also considerable intra-sample variability within the *L. lobatulus* population (e.g. 21 ‰ δ^{18}O and 18 ‰ δ^{13}C in the 140-145 cm sample; Table 6.2; Fig 6.7A, B). A least-squares regression of ^{14}C age vs depth, albeit with limited chronological control, indicates each 5 cm sample includes between 150 and 350 years of sedimentation. Therefore, the variability in the isotopic records is likely attributable to the homogenization of tests that formed under a range of environmental conditions. The total intra-sample isotopic variability would of course be greater than that shown, as any one test may be greater or less than the collective measure recorded from the ±5 tests analyzed.

6.3.2.2 *Lobatula lobatulus*

The δ^{18}O and δ^{13}C records for *L. lobatulus*, from the Walton Lake core, were plotted using a least-squares regression of ^{14}C age (years BP) vs depth (Fig. 6.8). Key changes within the record are highlighted and discussed separately.

Stage A (>9.3 - 8.8 ka BP) This stage comprises the diamict and immediately overlying sediments (300 - 260 cm). The diamict is constrained by a minimum age of 9270± 80 (TO-4757, Table 6.1). A date on shells of 9270±110 BP (GSC-3180; Blake 1981) from a 65 m terrace midway up the north arm of Makinson Inlet (Fig. 6.1) indicates a concurrent retreat of marine-based margins. The extremely depleted δ^{18}O compositions within this stage are considered to be consistent with dilution by glacial meltwater originating from an overlying glacial ice shelf. Based on measurements of the Wisconsinan-Holocene isotope step in the

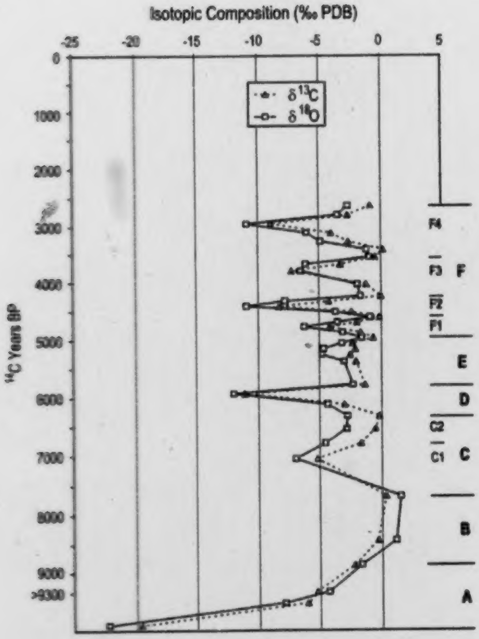

Figure 6.8 Walton Lake core stable isotope stratigraphy of *Lobatula lobatulus*, expressed in ^{14}C years BP. Isotope values are averages of replicate analyses. Major changes in the isotope record are subdivided and discussed within the text.

Agassiz and Devon Island Ice caps (Paterson *et al.* 1977; Koerner *et al.* 1987), the $\delta^{18}O_{SMOW}$ composition of late-Wisconsinan glacial ice from the Prince of Wales and Sydkap Ice caps would have been around -38 to -35 ‰. Isotopic dilution may also have been aided by the surface accretion of meteoric waters onto the postulated ice shelf. Assuming a 200 m thickening of the regional Ice Caps (Koerner *et al.* 1987), and the same meteorological (isobaric) controls as today (Fisher & Alt 1985), snow $\delta^{18}O_{SMOW}$ values around Hoved Island would have been between -45 and -38 ‰. These compare with present-day snow values of -28 ‰ reported for the Agassiz Ice Cap, and between -30 and -35 ‰ around Hoved Island (Koerner 1979). Local changes in the thickness and extent of the ice shelf, may be responsible for the variation seen within the replicate samples (Fig 6.7A, B). The trend to heavier $\delta^{18}O$ values is interpreted to reflect the thinning of the ice shelf, or possibly its initial breakup, permitting the entry of deeper, more saline waters.

The carbon isotope composition of marine carbonates, such as foraminifera, is considered to reflect the $\delta^{13}C$ of the DIC of the water in which they formed (Anderson & Arthur 1983). Marine carbonates forming in sea water, at isotopic equilibrium with the atmosphere, would have $\delta^{13}C$ values between 0 and 4‰ (Israelson *et al.* 1994). Negative $\delta^{13}C$ values are linked to the microbial respiration of carbon from organic matter and terrestrially-derived dissolved carbon (Mook & Vogel 1968, Sayles *et al.* 1988). The highly negative $\delta^{13}C$ values seen in this study (Figs. 6.7B and 6.8) are well beyond the range attributable to "vital effects" (Belanger *et al.* 1981; Aksu & Vilks 1988), or environmental factors such as sea ice cover and restricted CO_2 exchange between the atmosphere and water (de Vernal *et al.* 1992). Therefore, the $\delta^{13}C$ composition is considered to reflect dilution by terrestrially-derived organic carbon, of which there are likely two principal sources. Well-preserved, dead plant communities are exposed along the margins of many modern retreating glaciers within the High Arctic (Bergsma *et al.* 1984). This may also have been the case during deglaciation, where terrestrial organic carbon was transported to the sea, diagenetically altered and then incorporated into the carbonate of the foraminifera tests. However, the very low abundance of terrestrial organic detritus throughout most of the lower two-thirds of the core, suggests this is unlikely to be the principal cause of $\delta^{13}C$ depletion. The most likely source is the extensive outcrops of Devonian and Tertiary coal-bearing strata found at the mouth of Vendom Fiord, and along the south shore of inner and central Baumann Fiord (Ricketts & Embry 1984). Fine, detrital coal is found throughout the Walton Lake core in varying abundance, the diagenesis of which would likely produce the depleted $\delta^{13}C$ measurements seen here. The unusually strong correlation between the $\delta^{18}O$ and $\delta^{13}C$ records ($r^2 = 0.93$) reflects the dominance of the isotopic system within Baumann Fiord by glacial melt. Increased meltwater inputs recorded by the $\delta^{18}O$ stratigraphy are similarly reflected in the $\delta^{13}C$ compositions as a record of sediment (coal) input. Variations then in the $\delta^{13}C$ records should reflect differences in the meltwater/sediment source, as well as diagenesis. Coal outcrops most abundantly at the mouth of Vendom Fiord and along south-central Baumann Fiord (Ricketts & Embry 1984). Thus, meltwater from the Sydkap Ice Cap and its former northward extension would be expected to contribute the most coal-laden sediment.

Stage B (8.8 - 7.6 ka BP) Carbon and oxygen isotopic compositions attain values similar to those of modern-day Arctic water, with a possible slight enrichment in $\delta^{18}O$ (van Donk & Mathieu 1969; Tan & Strain 1980b; Bédard *et al.* 1981). This is interpreted as the result of the breakup and removal of the ice shelf by calving *en masse*. Significant melting of either the ice shelf, or regional ice caps is considered unlikely as this is not seen in the $\delta^{18}O$ record. Similarly, the absence of negative $\delta^{13}C$ values supports the notion of limited land-based retreat or melting of ice, as this would have delivered isotopically depleted coal detritus. Glacioisostatic studies around the fiord heads of west-central Ellesmere Island by Hodgson (1985) have shown this period (9 to 7.5 ka BP) to be characterised by limited postglacial uplift, therefore minimal reduction in ice load. England (1983, 1992, 1996) has demonstrated similar glacial histories and sea level adjustments along central and eastern Ellesmere Island. This pattern is explained by the breakup of mostly floating marine-based ice margins that resulted in minimal unloading. This slow rate of unloading was balanced by ongoing eustatic sea level rise, which produced a relatively stable sea level between 8.5 and 6.5 ka BP (England 1983, 1992). Subsequently, emergence, hence glacial unloading, was considered rapid, although the exact onset is uncertain.

Stage C (7.6 - 6.3 ka BP) The sudden negative excursion by both the carbon and oxygen stable isotopes may signify the initiation of land-based ice retreat, and renewed input of isotopically depleted meltwater (C1: 7030 BP, Fig. 6.8). Hodgson (1985) reports a date of 6980±90 BP (GSC-1957) on shells below topset beds at 63 m asl, 5 km upvalley from the head of Vendom Fiord. He concluded that retreat of ice from the head of Vendom fiord took place just prior to 7 ka BP.

The meltwater "pulse" of C1 (~7.3 to 6.7 ka BP), is followed by a period, C2 (6.7 to 6.3 ka BP), of only moderately depleted $\delta^{18}O$ values, while the $\delta^{13}C$ values rise to near modern compositions. This is interpreted as a period of general ice cap thinning, either by melting or drawdown, without substantial marginal retreat. Melting and land-based glacial retreat is unlikely to have occurred synchronously across the landscape. Ice masses centred on the Svendsen Peninsula and Braskeruds Plain occupied much lower elevations (350 - 550 m asl) compared to that of the Sydkap Ice Cap (350 - 1000 m asl) (Hodgson 1985), and would thus be more susceptible to a rise in equilibrium line altitude. Based on the outcrop pattern of coal-bearing strata within the region (Ricketts & Embry 1984), melting of glaciers on the Svendsen Peninsula and Braskeruds Plains would be expected to contribute the least amount of coal detritus, possibly explaining the low $\delta^{13}C$ depletion during this period (C2, Fig. 6.8).

Stage D (6.3 - 5.7 ka BP) Intense glacial melt and surface runoff during this period are indicated by the very negative carbon and oxygen isotope values. This is broadly conformable to the onset of rapid regional uplift reported by England (1990, 1992) and Hodgson (1985).

Without proper chronostratigraphic control on the glacioisostatic uplift of Hoved Island, it is not possible to accurately determine the paleo-water depth of Walton Lake. At the very least, however, it is known that the maximum depth of marine submergence decreased by at least 123 m (Holocene marine limit + depth of basin). The effect of this emergence on the isotopic record is that the depth of mixing, for any given input of isotopically dilute surface water, would become progressively more capable of producing the same isotopic fluctuations within the foraminifera. Another factor to consider is that the paleo-depth of the basin may actually be dictated by the position of the outlet, and not the bottom (27 m below). Theoretically then, the isotopic composition of the foraminifera at maximum marine submergence relates to water isotope characteristics at 96.5 m depth *vs.* 123.5 m. These depths are not outside the range of reported glacial meltwater isotopic dilution (Hillaire-Marcel 1981; Erlenkeuser 1985; Andrews *et al.* 1991; Israelson *et al.* 1994). The degree seen in this study, however, is considerably greater. This is likely attributable to the extremely negative isotopic composition of the glacial melt, and what must have been a relatively restricted circulation within Baumann Fiord.

Stage E (5.7 - 4.9 ka BP): A reasonably constant isotopic record indicates ongoing inputs of glacial meltwater, and coal-laden sediments throughout this period. Compared to Stage D, the $\delta^{18}O$ values are much less negative. This may suggest that the rapid retreat that occurred between 6.7 and 6 ka, resulted in a diminished areal extent of ice below the ELA, hence much reduced ablation. It may also have been augmented by a period of cooler summers.

It is uncertain whether low *L. lobatulus* test densities (Fig. 6.6) indicate an aversion to low salinity environments (no significant correlation exists between test densities and $\delta^{18}O$ values, $r^2 = 0.27$). Vilks & Deonarine (1988), working off Labrador, demonstrated that *L. lobatulus* has a wide salinity and temperature tolerance. Their measured range of salinity (28.6-34.8 ‰) would, however, have been less than that experienced at this site (~10-37 ‰).

Stage F (4.9 - 2.6 ka BP) This period is roughly accordant with a general cooling phase recognized throughout the High Arctic (Bradley 1990). It is not surprising then that many carbon and oxygen stable isotope values within this stage appear similar to modern-day Arctic water (van Donk & Mathieu 1969; Tan & Strain 1980b; Bédard *et al.* 1981). However, despite a regional cooling there have been at least four periods of enhanced glacial meltwater inputs (F1: 4.8 ka BP; F2: 4.4 ka BP; F3: 3.8 ka BP; F4: 2.9 ka BP). Intra-sample isotopic variability (Fig. 6.7A, B) is high during these periods, suggesting that the meltwater inputs were similarly variable. Summer melt from the Agassiz Ice Cap (core A84, Koerner & Fisher 1990) shows a sporadic pattern, suggesting that despite the general cooling trend, significant melting still occurred. The intensity of the isotopic dilution, and the temporal duration of the negative isotopic departures, was likely enhanced by the ongoing glacioisostatic uplift of the basin into the naturally more dilute surface waters. Insufficient numbers of *L. lobatulus* tests for analysis were found above 60 cm core depth.

6.3.2.3 *Portlandia arctica*

It is unknown why the stable carbon and oxygen isotope measures of the mollusc *P. arctica* are so discordant from those of the foraminifera *L. lobatulus* (Fig. 6.7C). Although few samples of *P. arctica* were run, an attempt was made to analyze those from periods representative of the complete *L. lobatulus* isotopic range. Replicate analyses of *L. lobatulus* from the same 5 cm core section demonstrate considerable isotopic variability, including periods approximating modern oceanic isotopic composition. It may simply be that *P. arctica* only formed shell carbonate during periods of higher salinity, hence, more "normal" isotopic composition. This selective behaviour, however, appears an unsatisfactory explanation as negative oxygen isotope values have been shown for *P. arctica* in the Champlain Sea sediments (Hillaire-Marcel 1981). Similarly, Aitken (pers. comm. 1995) reports finding living specimens in fiord waters off Baffin Island with salinity as low as 23‰. Redeposition by bottom currents of isotopically negative *L. lobatulus* tests from ice proximal areas to deeper ocean basins has been suggested by Erlenkeuser (1985). However, this also seems unlikely given the completeness of the record, the understanding of the tests' biology (Corliss 1985; Vilks & Deonarine 1988), and the absence of any evidence of sedimentological reworking.

The Hoved Island *P. arctica* isotopic data (using interpreted ^{14}C ages, Fig. 6.8) and the ^{13}C data of *P. arctica* samples dated by the Geological Survey of Canada (GSC), were plotted against that of Andrews *et al.* (1993 [their Figure3]; Fig. 6.9). While there is a weak agreement between the δ^{18}O records, the δ^{13}C values of *P. arctica* samples show little similarity to those of Andrews et al. (1993). Andrews *et al.* (1993) interpreted their data as reflecting strong meltwater inputs between 11 and 9 ka BP. It is puzzling though why shells in ice proximal locations such as theirs were not more negative, especially considering the enormous volumes of meltwater the Laurentide Ice Sheet must have generated. Meltwater

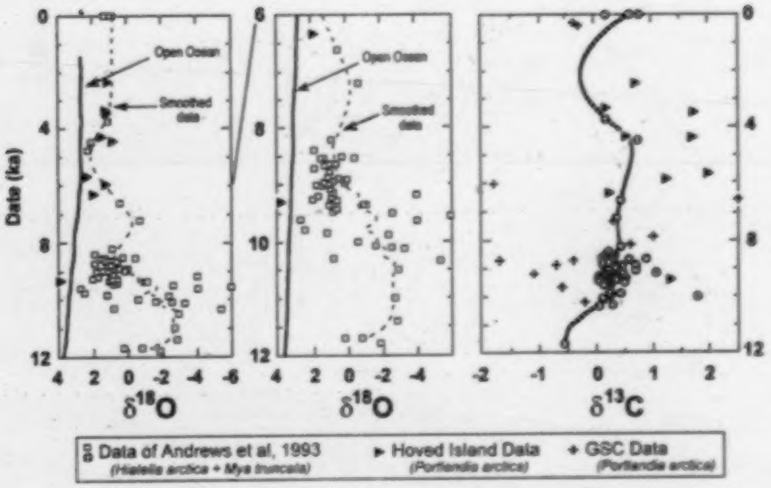

Figure 6.9 (*after*, Andrews *et al.*, in Geology, v.21, p.881, 1993, the Geological Society of America. Reproduced and modified with permission). Isotopic data from the Walton Lake *Portlandia arctica* samples, as well as *P. arctica* samples from the Geological Survey of Canada (GSC) radiocarbon data base. Data from Andrews *et al.* (1993), is from radiocarbon dated samples of the near-surface marine mollusks, *Mya truncata* and *Hiatella arctica* collected from raised marine deposits on the Canadian Arctic mainland and islands. GSC collections of *P. arctica* are almost exclusively from Ellesmere Island.

discharges from local ice caps within the Baumann Fiord region would have been considerably less. Even despite isotopic differences in meteoric waters between the two regions, the dilution factor would presumably have been greater along the margin of the Laurentide Ice Sheet.

6.4 CONCLUSIONS

This study has demonstrated intervals of isotopically depleted ($\delta^{18}O$) glacial meltwater inputs throughout the Holocene, within the Baumann Fiord region of Ellesmere Island, as recorded by the epibenthic foraminifera, *Lobatula lobatulus*. These intervals compare favourably with the proposed rates of glacial retreat published previously based on dated ice margins and rates of postglacial rebound (England 1983, 1990, 1992; Hodgson 1985; Bednarski 1995; Bell 1996). Furthermore, the stable isotope record provides new insights into the rate of ice retreat during the early to mid-Holocene, and likely demonstrates details of a stepped ice retreat that have been beyond the resolution provided by studies of continuous postglacial emergence. Ice retreated rapidly between 7.6 and 7.0 ka BP (Stage C1; Fig. 6.8) and 6.3 to 5.7 ka BP (Stage D), but was interrupted by intervals of slow retreat, 8.8 to 7.6 ka BP (Stage B) and 7.0 to 6.3 ka BP (Stage C2). After 5.7 ka BP, regional cooling ensued, however, this was punctuated by intervals of increased melting around 4.8, 4.4, 3.7 and 2.8 ka BP (F1, F2, F3 and F4, respectively).

The strong correlation (r^2=0.93) between the $\delta^{13}C$ and $\delta^{18}O$ data reflects the dominance of the overall hydrological and isotopic system by the regional glaciers. Isotopic variations are considerably greater than those attributable to known vital effects, or those previously recorded in circumpolar benthic studies. This is likely the result of many factors, including: the extremely light $\delta^{18}O$ composition of regional meteoric waters, the diagenesis of coal-rich sediments, an overall restricted circulation pattern within Baumann Fiord, and the unique environment of the isolation basin that becomes progressively uplifted over time. Reasons for the incongruity between the isotopic composition of the mollusc, *Portlandia arctica*, and the foraminifer, *L. lobatulus*, are unknown but may be resolved through studies of the modern environment.

A tentative reconstruction of the glacial history of the region is proposed. Prior to 9.3 ka BP, a grounded glacier occupying central Baumann Fiord became decoupled from its bed, forming a glacial ice shelf. Retreat of the ice shelf occurred between ~9.3 and 8.8 ka BP. The absence of any isotopic dilution during this phase suggests that it broke up by calving *en masse*, with little melting. Similarly, there must have been little melting of the regional ice caps during this period which is consistent with the limited regional emergence (unloading) observed for this area (England 1983, 1992). Large-scale melting of the regional ice caps occurred post 7.6 ka BP, centred on ~7.3 and 6.0 ka BP. This also coincided with the initiation of land-based retreat as interpreted from the negative carbon isotope values considered to reflect the inwash of terrestrial carbon (coal). However, this was interrupted by an interval of slow melt between 7.0 and 6.3 ka BP when there was only minor oxygen isotope dilution and no significant carbon isotope dilution. This is interpreted to record limited marginal retreat, again consistent with the observed slow regional emergence. Between 6.3 and 5.7 ka BP, the first major meltwater "pulse" occurred, which coincides with the onset of rapid emergence reported to the north (England 1990). Meltwater inputs were strongly reduced between 5.7 and 4.9 ka BP, suggesting an important interval of glacier stability. Relative volumes of meltwater, however, are difficult to assess throughout the mid- to late-Holocene, because the Walton Lake basin would have been slowly uplifted until it finally emerged from the sea around 900 years ago. Thus, the same isotopic dilution over time would require less input of glacial meltwater. Nevertheless, throughout the period 4.9 to 2.6 ka BP, there were at least four periods of meltwater input of seemingly increasing duration, recording intermittent ice retreat within an interval of recognized cooling (Bradley 1990). The chronology and nature of the events proposed here provides new insights and increased resolution pertaining to ongoing glacial geomorphological investigations and the history of Holocene paleoenvironmental change within the Canadian High Arctic.

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SUMMARY

The existence of unglaciated terrain beyond the limit of ice during the Last Glacial Maximum (LGM) was not proven in either the Lake Hazen or Hoved Island regions of Ellesmere Island, Canadian High Arctic. Within the Lake Hazen region, lateral meltwater channels, moraines, buried glacial ice and the cosmogenic ^{36}Cl dating of erratics and bedrock record the complete inundation of the landscape by ice during the LGM. This thus confirms the hypothesis of England (1998, in press). Coring of extant lake basins beyond the formerly proposed limit of ice during the LGM (England 1978, 1983) also supports a regional ice cover, as AMS radiocarbon dates of organic remains are all younger than 7.6 ka BP.

On Hoved Island, coring of three emergent lake basins also failed to document any pre-glacial sediments. Recent glacial reconstructions by Bednarski (in press), Dyke (in press) and Ó Cofaigh (1998) propose that during the LGM, glaciers inundated western Ellesmere Island, eastern Axel Heiberg Island and Devon Island, including the intervening marine channels. Thus the ^{14}C date of 9270±80 BP (TO-4757) from near the base of the Killam Lake core (Hoved Island), provides a minimum estimate for deglaciation of Hoved Island and central Baumann Fiord.

The question of refugia within the Lake Hazen region remains enigmatic. Both the diversity and abundance of spiders, bryophytes, and ostracodes have been cited by different authors as indication that refugia may have existed within this region during the LGM (Leech 1966; Brassard 1971; Røen 1981). However, given the glacial history proposed by England (1998, in press) and that documented in this study, only small isolated nunataks may have existed in the Grant Land Mountains and on northern Judge Daly Promontory. These are unlikely to have sustained these flora and fauna, and therefore if refugia existed, they must have lain elsewhere. With regards to the ostracodes, there appears to be a distinct separation in the populations of northern Greenland (to which the Hazen samples are most similar), and those in western and southern Greenland (Røen 1981). Perhaps northern Greenland maintained refugia during the LGM, from which species migrated to the Lake Hazen region. This remains to be tested.

During the LGM, a large trunk glacier emanating from the Grant Land Mountains flowed northeast across Lake Hazen Basin and Hazen Plateau, coalescing with Agassiz ice flowing northeast down the Dodge River valley and Archer Fiord. The Ellesmere ice was abutted by Greenland ice blocking the outlet of Lady Franklin Bay, causing the Agassiz ice to override outer Hazen Plateau along a coalescent Ellesmere-Greenland ice boundary. This onshore flow is recorded by erratics above Eastwind Bay and striae along Turnabout River. The actual position of the coalescent boundary is unknown, though has been indicated by England (1998, in press) to lie along the westernmost limit of granite (Greenland) erratics. This remains to be tested by cosmogenic dating techniques. The absence of granite erratics above Holocene marine limit within central and inner Lady Franklin Bay, argues that Ellesmere ice must have blocked entry of Greenland ice. Therefore, once Greenland ice had impinged upon outer Hazen Plateau, there must have been a rapid buildup of Ellesmere ice in the adjacent fiords. This suggests that when Greenland ice was advancing onto outer Hazen Plateau, Grant Land Mountain and Agassiz ice had already infilled Lake Hazen Basin and advanced to the fiord heads. Maximum estimates for growth of ice up to the LGM are provided by dates as young as 23 ka BP on shells in till along northern Nares Strait (England 1998, in press), and a date of 24 ka BP on a *Salix arctica* twig in deglacial sediments at the head of Clements Markham Inlet (Bednarski 1986).

Deglaciation of northeastern-most Ellesmere Island occurred by 10.1 ka BP, though trunk ice continued to occupy Robeson Channel, blocking entry of the sea (England 1978; 1998; in press). By 9.6 ka BP, central Hall Land (Greenland) was deglaciated (England 1985), implying a substantial retreat and thinning of the trunk ice in Robeson Channel. In contrast to the 9 ka BP glacial reconstruction proposed by England which maintains a coalescent Ellesmere-Greenland ice boundary ~15 km inland, this study hypothesizes that at 9 ka BP the trunk ice was confined within Robeson Channel, abutting the Ellesmere Island coast. Separation between trunk ice in Robeson Channel and Grant Land Mountain ice occupying outer Hazen Plateau (cf. Retelle 1986) would have occurred between 10 and 9 ka BP. Kames, kame terraces and deltas formed along these separating margins were previously interpreted to record the retreat of Greenland ice off the Ellesmere coast (Retelle 1986). This study, however, proposes that it was Agassiz ice which formed the western, coastal margin of the trunk ice in Robeson Channel. This requires that the granite erratics within some of the deltas, kames and lateral meltwater channels associated with the retreat of the trunk ice, be redeposited by Ellesmere (Agassiz) ice, overriding Greenland till. Testing of this hypothesis may be carried out by inspecting these same features to see if they contain any diagnostic Ellesmere (Agassiz and Grant Land Mountain) erratics.

Between 9 and 8 ka BP ice had retreated to the heads of many of the regional fiords, recorded by ice-contact marine limit deltas. Moraines linked to these deltas, the Hazen Moraines, were previously considered to mark the limit of ice during the last glaciation (England 1978, 1983). However, given the recent glacial reconstructions proposed by (England 1998, in press), and results of this study, it is now recognized that the Hazen Moraines do not mark a chronostratigraphic boundary. Instead they represent the land-based stabilized margins of formerly marine-based glaciers, differences in the age of which reflect local bathymetry, topography and flow dynamics. The Craig Lake Moraine, previously considered part of the Hazen Moraines (England 1978, 1983), is constrained by a peat date of ~6.0 ka BP, indicating that it is not coeval with the deposition of regional marine limit deltas and associated moraines.

Also between 9 and 8 ka BP, flow of Grant Land Mountain trunk ice occupying Lake Hazen Basin and Hazen Plateau shifted southwards, and outlet glaciers became confined within large valleys. *En echelon* lateral meltwater channels record the thinning of the trunk ice. The fact that these channels do not extend to the summits of the local highlands implies that during retreat, separation occurred between the trunk ice and plateau ice caps left occupying the summits. Meltwater channels that cross-cut those of the retreating trunk ice point to the subsequent growth of plateau ice, some time prior to 7.6 ka BP, at which point the plateau ice also had begun retreating.

At 7 ka BP, ice margins within Lake Hazen Basin appear to have stabilized, retreating only slightly between then and 6 ka BP, when the extensive Craig Lake Moraine was established. This relative stability is thought to reflect an entirely land-based ice mass, and the westward retreat of ice from central Hazen Plateau, where previously it had impounded large proglacial lakes. Diatom proxy records of lake ice cover from Brainard and Hart lakes on Hazen Plateau confirm that this period remained cold. A date of 5.3 ka BP on mosses from coarse sorted sediments in Connell Lake indicates deposition in a proglacial lake, requiring that trunk ice continued to occupy the Ruggles River valley. Final retreat of ice within proto-Lake Hazen occurred by 5.0 ka BP, by which point ice appears to have retreated to near its modern limits throughout the study area.

The final retreat of ice from Lake Hazen Basin heralds the amelioration of the regional climate. Gradual increases in diatom abundance between 5 and 4 ka BP in each of the four study lakes suggest progressive decreases in summer lake ice cover. Peak diatom abundance occurs between 4 and 3 ka BP, indicating this was the warmest period of the Holocene recorded in this study. However, other studies throughout the Canadian High Arctic indicate a dramatic warming during the past century (Douglas *et al.* 1994; Overpeck et al. 1997). While similar changes in diatom species diversity in the uppermost sediments of Appleby Lake appear similar to the results of Douglas *et al.* (1994), uncertainties regarding preservation of the uppermost sediments do not allow a useful comparison to be made. At 3 ka BP, diatom abundances drop in Brainard Lake, but remain high in Appleby Lake until ~1.9 ka BP, after which they drop. This is interpreted to reflect a gradual cooling and lowering of the regional snowline, such that lakes higher on the plateau (i.e., Brainard) experienced a more pervasive summer lake ice cover than lakes lower down in Lake Hazen Basin (i.e. Appleby). The decline in diatom abundance within Appleby Lake following 1.9 ka BP indicates a further cooling, reaching its lowest point ~1.5 ka BP, after which, conditions appear to warm.

The paleoenvironmental reconstructions proposed here are not in complete agreement with those from elsewhere in the Canadian High Arctic. Reasons for the disparity include the unique topoclimatic setting and continentality of the Lake Hazen region, as well as the bias of other paleoenvironmental records from coastal areas, and the possible effects that local marine conditions may have had upon them.

On Hoved Island, extremely depleted stable isotope records of *Lobatula lobatulus* within the diamict of Walton Lake, and an exponential rise in foraminifera abundance, support the interpretation that the diamict formed by rain-out from an ice shelf ~9.3 ka BP.

Continued break up and retreat of ice within central Baumann Fiord appear to have occurred largely by calving with little melting, as there is no indication of isotopic depletion between 9.3 and 8.8 ka BP. Shifts in the isotopic records to more negative values around 7.6 and 6 ka BP indicate strong glacial meltwater inputs, coincident with the initiation of land-based retreat (cf. Hodgson 1985). Meltwater inputs were reduced between 5.7 and 4.9 ka BP, suggesting an important interval of glacial stability. Subsquently (4.9 - 2.6 ka BP) a series of meltwater "pulses" recorded intermittent ice retreat within an interval of recognized cooling (Bradley 1990).

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